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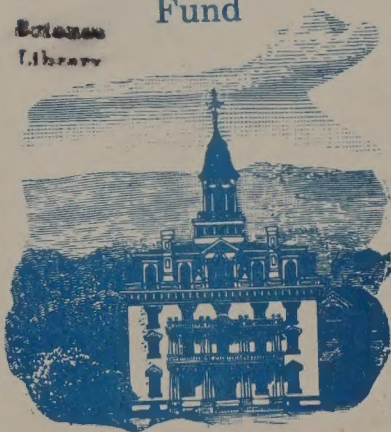
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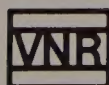
PETROLOGY/*D. R. Bowes*

ENCYCLOPEDIA OF EARTH SCIENCES

The
ENCYCLOPEDIA
of
IGNEOUS AND
METAMORPHIC
PETROLOGY

EDITED BY

D. R. Bowes
Department of Geology
University of Glasgow



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Van Nostrand Reinhold
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PREFACE

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The concept of Rhodes W. Fairbridge of a series of reference works, not only for earth scientists, but also for those of other disciplines wishing ready access to information about the earth sciences, has resulted in the *Encyclopedia of Earth Sciences*. This now fulfils an important function in the libraries of both institutions and individuals. As was the case for previous volumes, contributors to this volume have been asked to include not only background information for those not expert in the particular field, but also data and a bibliography that permit the various aspects of the topic to be followed up. The overall aim has been to compile a volume acceptable to experts, such as the contributors themselves, but that makes information readily accessible to earth scientists with other backgrounds, scientists from other disciplines and the large body of people outwith the scientific professions who are interested in the Earth.

Petrologists working in the field of igneous petrology particularly, and in metamorphic petrology also, know of the huge vocabulary of rock names, many of which are variants of common rock assemblages that are given local names. There is little doubt that such proliferation of terms has been a major factor inhibiting communication in the earth sciences. It is hoped that the extensive index together with sections on **Other terms** incorporated in entries about major rock types, will assist in making information about the less well known, obscure and even obsolete terms that occur in literature, readily accessible. In no way, however, should the inclusion of a term in this volume be taken as an assessment of its value, or an endorsement of its continued use. The *AGI Glossary* indicates those terms considered

to be obsolete and gives recommendations concerning continued use, or otherwise, of terms, although that is obscure to some earth scientists in one part of the world is commonly used by other earth scientists in other parts of the world.

More than one hundred contributors from eighteen countries have contributed to this volume. Such an international representation is essential for the best possible presentation, but it also brings out the varying usage and even spelling of terms. There will be no earth scientist who makes reference to this volume who will find that everything corresponds with his or her own usage. However it is hoped that the various backgrounds represented will assist in making the enormous range of earth science literature more easily accessible and more intelligible to scientists worldwide.

Integrating the various contributions into a unified whole has been a major task, and the help of my wife in this has been invaluable as has been her patience in proof reading and indexing. Many articles have greatly benefited from the expertise of Dr Adrian F. Park who, while at the University of Glasgow, undertook the major task of adding much recent information. Others at the Department of Geology, University of Glasgow, who have made major contributions to the compiling of this work are Professor Bernard E. Leake, Mrs Sheila Hall (drafting), Mr Douglas Maclean (photography) and Mrs Mary Fortune, who with great expertise and patience has typed, and retyped, major sections of the volume and assisted in the voluminous correspondence and day to day administration.

Mr Alan Rodwell of Universities Press (Belfast) Ltd and Ms Stefania Talflinska of

PREFACE

Van Nostrand Reinhold, and those they represent, are sincerely thanked for their efforts and encouragement during the production of this volume. And Professor Rhodes W. Fairbridge who began the work on this volume some considerable time ago is thanked, not only for his contribution to this particular volume, but for his major contribution to the earth sciences repre-

sented by the whole encyclopedia series. But the major credit line must go to the contributors, many of whom have exercised patience they did not know existed, and all of whom have drawn on their wealth of professional expertise in their contributions.

D. R BOWES

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MAIN ENTRIES

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 Volcanic breccia
 Volcanic glass
 Volcanic neck
 Volcanic rocks
 Volcanoes and volcanism

 Water in rock melts

 Xenolith

 Zeolite facies

CONTRIBUTORS

INA B. ALTERMAN, National Research Council, 2101 Constitution Avenue, Washington D.C. 20418, U.S.A. *Phyllite; Slate*

M. P. ATHERTON, Department of Earth Sciences, University of Liverpool, P.O. Box 147, Liverpool L69 3BX, England. *Burial metamorphism; Depth zones; Metamorphism—controls.*

D. K. BAILEY, Department of Geology, University of Reading, Whiteknights, Reading RG6 2AB, England. *Alkaline rocks—undersaturated; Peralkaline igneous rocks.*

R. A. BAILEY, U.S. Geological Survey, 345 Middlefield Road, Menlo Park, California 94025, U.S.A. *Volcanic glass.*

Y. K. BENTOR, Scripps Institute of Oceanography, University of California—San Diego, La Jolla, California 92093, U.S.A. *Granodiorite.*

A. C. BISHOP, Department of Mineralogy, British Museum (Natural History), Cromwell Road, London SW7 5BD, England. *Aplite; Greisen.*

G. BODVARSSON, School of Oceanography, Oregon State University, Corvallis, Oregon 97331, U.S.A. *Fumarole; Geyser; Hot ground; Hot spring; Solfatara; Thermal activity.*

D. R. BOWES, Department of Geology and Applied Geology, University of Glasgow, Glasgow G12 8QQ, Scotland. *Abbreviations; Alkaline rocks—undersaturated: other petrological terms; Alteration, replacement; Amphibolite; Analcimite; Appinite; Aureole; Diabase; Experimental petrology; Häüynite; Igneous petrology; Igneous petrology—terms; Igneous rocks—other terms; Injection complex; Island arcs—petrology; Latite; Magmatic—tectonic*

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JOHN BRADLEY, deceased. *Dyke (dike); Laccolith; Sill.*

J. ROBERT BUTLER, Department of Geology, University of North Carolina, Chapel Hill, North Carolina 27599–3315, U.S.A. *Facies of thermal metamorphism; Pyrometamorphism; Pyrometasomatism.*

BRIAN CHADWICK, Department of Geology, University of Exeter, North Park Road, Exeter EX4 4QE, Devon, England. *Contact metasomatism; Petrographic microscope.*

G. A. CHALLIS, New Zealand Geological Survey, P.O. Box 30368, Lower Hutt, New Zealand. *Harzburgite.*

D. W. CHIPMAN, Lamont–Doherty Geological Observatory of Columbia University, Palisades, New York 10964, U.S.A. *Retrograde metamorphism.*

NIKOLAS I. CHRISTENSEN, Department of Earth and Atmospheric Sciences, Purdue University, West Lafayette, Indiana 47907, U.S.A. *Oceanic crust.*

K. COE, Department of Geology, University of Exeter, North Park Road, Exeter EX4 4QE, Devon, England. *Crypto-volcanic structure.*

J. W. COLE, Department of Geology, University of Canterbury, Private Bag, Christchurch, New Zealand. *Tephra*.

K. C. CONDIE, Department of Geoscience, New Mexico Institute of Mining and Technology, Socorro, New Mexico 87801, U.S.A. *Igneous rocks—tectonic setting*.

DOUGLAS S. COOMBS, Department of Geology, University of Otago, P.O. Box 56, Dunedin, New Zealand. *Brownstone facies; Lawsonite–albite–chlorite facies; Prehnite–pumpellyite facies; Pumpellyite–actinolite schist facies; Zeolite facies*.

K. G. COX, Department of Earth Sciences, Oxford University, Parks Road, Oxford OX1 3PR, England. *Volcanic rocks*.

S. J. CUTHBERT, Department of Geology and Applied Geology, University of Glasgow, Glasgow G12 8QQ, Scotland. *Reaction rim*.

P. E. DAMON, Laboratory of Isotope Geochemistry, Department of Geosciences, University of Arizona, Tucson, Arizona 85721, U.S.A. *Silicate melting*.

B. DASH, Department of Geology, Utkal University, Bhubaneswar, Orissa 751 004, India. *Khondalite*.

J. B. DAWSON, Department of Geology, University of Sheffield, Mappin Street, Sheffield S1 3JD, England. *Kimberlite*.

T. DEANS, 20 Grove Park Road, Chiswick, London W4 3SD, England. *Carbonatite*.

M. R. DENCE, The Royal Society of Canada, 344 Wellington Street, Ottawa, Ontario K1A 0N4, Canada. *Shock metamorphism*.

E. DEN TEX, Rijksmuseum van Geologie en Mineralogie, Garenmarkt 1B, 2311 PG Leiden, The Netherlands. *Steinmann Trinity*.

D. DE WAARD, Rt 2, Box 53A, Greenville, Florida 32331, U.S.A. *Diapirism*.

W. R. DICKINSON, Department of Geosciences, University of Arizona, Tucson, Arizona 85721, U.S.A. *Geosynclinal volcanism; Pacific and Atlantic suites*.

J. DIDIER, Département de Géologie et Minéralogie, Université Blaise Pascal, 5 Rue Kessler, 63038 Clermont-Ferrand, France. *Enclave*.

R. V. DIETRICH, Department of Geology, Central Michigan University, Mount Pleasant, Michigan 48858, U.S.A. *Phacolith; Ptygmatic vein*.

G. P. DURANT, Hunterian Museum, University of Glasgow, Glasgow G12 8QQ, Scotland. *British Tertiary Volcanic Province; Rhyolite*.

I. W. FERGUSON, Digital Equipment Corporation, Mosshill Industrial Estate, Ayr KA6 6BE, Scotland. *Double-diffusive convection; Magmatic crystallization in open systems*.

E. H. FRANCIS, Department of Earth Sciences, University of Leeds, Leeds LS2 9JT, England. *Tuffsite; Volcanic neck*.

IAN C. FREESTONE, British Museum Research Laboratory, Great Russell Street, London WC1B 3DG, England. *Liquid immiscibility*.

W. S. FYFE, Department of Geology, University of Western Ontario, London, Ontario N6A 3K7, Canada. *Metamorphic reactions; Neutrality—pH and temperature variation*.

A. K. GALWEY, Department of Chemistry, The Queen's University, Belfast BT7 5AG, Northern Ireland. *Nucleation of metamorphic minerals; Reaction rates in metamorphism*.

ANNA T. GAVASCI, 2001 Kimbrough Green, Germantown, Tennessee 38138, U.S.A. *Mafic and ultramafic inclusions in igneous rocks; Textures of igneous rocks*.

A. M. GOODWIN, Department of Geology, University of Toronto, Toronto M5S 1A1, Canada. *Archaean volcanism; Greenstone.*

D. H. GREEN, Department of Geology, University of Tasmania, Hobart, Tasmania 7001, Australia. *Pyrolite; Ultramafic belts; Upper mantle—experimental petrology.*

C. D. GRIBBLE, Department of Geology and Applied Geology, University of Glasgow, Glasgow G12 8QQ, Scotland. *Assimilation; Hybridization; Reaction principle and reaction series.*

J. V. GUY-BRAY, Mineral Resources Branch, Natural Resources and Energy Division, Department of Technical Cooperation for Development, United Nations, New York, New York 10017, U.S.A. *Lopolith.*

RICHARD D. HAGNI, Department of Geology and Geophysics, University of Missouri—Rolla, Rolla, Missouri 65401, U.S.A. *Igneous rocks—classification and nomenclature.*

N. M. HALDEN, Department of Geological Sciences, University of Manitoba, Winnipeg R3T 2N2, Canada. *Element partitioning in petrogenesis; Geochemical modeling in petrogenesis; Migmatite—origin of neosome.*

A. HALL, Department of Geology, Royal Holloway and Bedford New College, Egham Hill, Egham, Surrey TW20 0EX, England. *Trend surface analysis in petrology.*

D. L. HAMILTON, Department of Geology, University of Manchester, Manchester M13 9PL, England. *Water in rock melts.*

A. L. HARRIS, Department of Earth Sciences, University of Liverpool, P.O. Box 147, Liverpool L69 3BX, England. *Polyphase metamorphic mineral growth.*

P. G. HARRIS, Department of Geology, University of Western Australia, Nedlands 6009, Western Australia. *Core.*

PAUL HENDERSON, Department of Mineralogy, British Museum (Natural His-

tory), Cromwell Road, London SW7 5BD, England. *Nucleation of minerals in magmas.*

A. HERRIOT, Department of Geology and Applied Geology, University of Glasgow, Glasgow G12 8QQ, Scotland, *Lamprophyre.*

HEINRICH D. HOLLAND, Hoffman Laboratory, Harvard University, 20 Oxford Street, Cambridge, Massachusetts 02138, U.S.A. *Hydrothermal processes.*

F. V. HOLUB, Department of Petrology, Charles University, Albertov 6, 128 43 Prague 2, Czechoslovakia. *Basalt; Calc-alkaline rocks; Durbachite.*

A. M. HOPGOOD, Department of Geology, University of St Andrews, St Andrews, Fife KY16 9ST, Scotland. *Migmatite—structural relationships.*

ROBERT HUTCHISON, Department of Mineralogy, British Museum (Natural History), Cromwell Road, London SW7 5BD, England. *Meteorites—petrology.*

RICHARD H. JAHNS, deceased. *Pegmatite.*

E. JELÍNEK, Department of Mineralogy and Geochemistry, Charles University, Albertov 6, 128 43 Prague 2, Czechoslovakia. *Basalt; Calc-alkaline rocks.*

K. A. JONES, Department of Geology, The Queen's University, Belfast BT7 1NN, Northern Ireland. *Nucleation of metamorphic minerals; Reaction rates in metamorphism.*

G. A. JOPLIN, Research School of Earth Sciences, The Australian National University, Canberra, ACT 2600, Australia. *Hypabyssal rocks.*

M. B. KATZ, Department of Applied Geology, University of New South Wales, P.O. Box 1, Kensington, N.S.W. 2033, Australia. *Gneiss.*

H. KLÁPOVÁ, Department of Petrology, Charles University, Albertov 6, 128 43

Prague 2, Czechoslovakia. *Basalt; Calc-alkaline rocks; Petrographic province.*

R. KRETZ, Department of Geology, University of Ottawa, Ottawa, Ontario K1N 6N5, Canada. *Metamorphic mineral growth.*

M. A. LAPPIN, Department of Geology and Petroleum Geology, Marischal College, University of Aberdeen, Aberdeen AB9 1AS, Scotland. *Granulite facies.*

W. R. LAUDER, deceased. *Harzburgite.*

BERNARD E. LEAKE, Department of Geology and Applied Geology, University of Glasgow, Glasgow G12 8QQ, Scotland. *Petrochemical calculations; Statistics in petrology.*

D. J. LEVESON, Department of Geology, Brooklyn College of the City University of New York, Brooklyn, New York 11210, U.S.A. *Orbicular rocks.*

E. LOCARDI, Centro Ricerche Energia Casaccia, S.P. Anguillarese 301, Casella Portale n. 2400, 00100 Roma A.D., Italy. *Mediterranean suite.*

S. MAALØE, Geologisk Institut, Universitetet i Bergen, Allegaten 41, 5007 Bergen, Norway. *Magmatic currents and flow.*

ALEXANDER R. McBIRNEY, Center for Volcanology, University of Oregon, Eugene, Oregon 97403, U.S.A. *Andesite and dacite; Xenolith.*

J. G. MacDONALD, Department of Adult and Continuing Education, University of Glasgow, Glasgow G12 8QQ, Scotland. *Cone sheets; Ring dyke.*

IAN McDOUGALL, Research School of Earth Sciences, The Australian National University, GPO Box 4, Canberra, ACT 2600, Australia. *Granophyre.*

DUNCAN McKIE, Department of Earth Sciences, Downing Street, Cambridge CB2 3EQ, England. *Fenite.*

ERIC A. K. MIDDLEMOST, Department of Geology and Geophysics, University of

Sydney, Sydney, N.S.W. 2006, Australia. *Magmatic differentiation.*

T. MORI, Department of Geology and Mineralogy, Kyoto University, Kyoto, 606 Japan. *Upper mantle—experimental petrology.*

M. MUNRO, Department of Geology and Petroleum Geology, Marischal College, University of Aberdeen, Aberdeen AB9 1AS, Scotland. *Banding in igneous rocks; Igneous intrusions—structural behavior.*

CHERUKUPALLI E. NEHRU, Department of Geology, Brooklyn College of the City University of New York, Brooklyn, New York 11210, U.S.A. *Petrographic methods.*

BERT E. NORDLIE, Department of Earth Sciences, Iowa State University, Ames, Iowa 50011, U.S.A. *Eutectic melting; Hornfels; Magmatic crystallization and paragenesis; Volatiles.*

C. OFTEDAHL, deceased. *Sövite.*

V. M. OVERSBY, Chemistry and Materials Science Department L-310, Lawrence Livermore National Laboratory, Livermore, California 94550, U.S.A. *Phase rule.*

ADRIAN F. PARK, Department of Geology, University of New Brunswick, Fredericton, New Brunswick E3B 5A3, Canada. *Ariégite; Astrobleme; Batholith; Cataclastic rocks; Charnockite; Charnockite suite; Contact (thermal) metamorphism; Deserpentinization; Flood basalt; Glaucophane schist facies; Granite; Granites and mineralization; Intrusion—layered; Lherzolite; Mid-ocean ridge and ocean-floor petrology; Ocean-island igneous rocks; Ophiolite; Peridotite; Plutonic rocks—classification and nomenclature; Pyroxenite; Serpentinite; Serpentinization; Skarn; Spinifex texture; Tektite; Ultramafic lavas; Volcanoes and volcanism.*

IAN PARSONS, Department of Geology and Geophysics, University of Edinburgh, Kings Buildings, Edinburgh EH9 2JW, Scotland. *Diorite and tonalite; Feldspath-*

oidal rocks—genesis; Intermediate igneous rocks; Monzonite (syenodiorite); Nephelinite; Phonolite; Syenite; Trachyte.

R. T. PRIDER, Department of Geology, University of Western Australia, Nedlands 6009, Western Australia. *Augitite, Lamproite; Leucitite; Limburgite; Melilitite.*

W. J. REA, Department of Education and Science, Government Buildings, Marston Road, Oxford OX3 0YT, England. *Petrology—history.*

DORIS L. REYNOLDS, deceased. *Fluidization.*

P. B. ROBERTSON, Geological Survey of Canada, Energy Mines and Resources, 1 Observatory Crescent, Ottawa, Ontario K1A 0Y3, Canada. *Shock metamorphism.*

J. A. RODDICK, Geological Survey of Canada, 100 W Pender Street, Vancouver, British Columbia V6B 1R8, Canada. *Circum-Pacific plutonic and volcanic activity.*

E. P. SAGGERSON, Department of Geology and Applied Geology, University of Natal, King George V Avenue, Durban, Natal 4001, Republic of South Africa. *Regional metamorphism.*

C. K. SEYFERT, Department of Geosciences, State University of New York, Buffalo, New York 14222, U.S.A. *Granites and volcanism.*

A. E. SHIMRON, Geological Survey of Israel, 30 Malkhei Ysrael, Jerusalem 95501, Israel. *Schist.*

C. H. SIMONDS, SN7, Life Support Development, Lockheed Missiles & Space Co. Inc., 1150 Gemini Blvd, Houston, Texas 77058, U.S.A. *Lunar petrology.*

JOHN A. T. SMELLIE, Swedish Geological Company, Box 1424, 751 44 Uppsala, Sweden. *Atoll garnet.*

K. SMULIKOWSKI, deceased. *Eclogite.*

J. SOUČEK, Department of Petrology, Charles University, Albertov 6, 128 43 Prague 2, Czechoslovakia. *Basalt; Calc-alkaline rocks.*

ALAN SPRY, Alan H. Spry & Associates, 512 Kensington Road, Wattle Park 5066, South Australia. *Buchite; Columnar jointing; Hornfelsic textures; Phenocryst; Porphyroblast; Porphyroclast; Porphyry; Snowball garnet.*

M. ŠTEMPROK, Geological Survey, Malostranské nám. 19, 118 21 Prague 1, Czechoslovakia. *Pegmatitic ore deposits.*

J. TARNEY, Department of Geology, University of Leicester, Leicester LE1 7RH, England. *Magma; Magmatic crystallization in metamorphic environments; Picrite.*

TARO TAKAHASHI, Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York 10964, U.S.A. *Phase transformation.*

T. P. THAYER, 3211 W Maritana Drive, St Petersburg, Florida 33706, U.S.A. *Alpine mafic magma stem; Ultramafic rocks.*

CHARLES P. THORNTON, Department of Geosciences, Pennsylvania State University, University Park, Pennsylvania 16802, U.S.A. *Calderas and volcano-tectonic depressions; Differentiation index; Lava and lava eruptions; Pyroclastic eruptions.*

PETER J. TRELOAR, Department of Geology, Imperial College, Prince Consort Road, London SW7 2BP, England. *Metamorphic petrology—use of thermodynamics.*

W. E. TREMLETT, Department of Geology and Applied Geology, University of Glasgow, Glasgow G12 8QQ, Scotland. *Nuée ardente; Pyroclastic flows.*

T. G. VALLANCE, 3 Kelburn Road, Roseville, N.S.W. 2069, Australia. *Spilite.*

W. J. VERWOERD, Department of Geology, University of Stellenbosch, Stellenbosch 7600, Republic of South Africa. *Palaeovolcanology*.

L. VILLARI, Istituto Internazionale di Vulcanologia, Viale Regina Margherita 6, 95123 Cantania, Italy. *Endogenous dome*.

GERHARD VON GRUENEWALDT, Department of Geology, University of Pretoria, Pretoria 0002, Republic of South Africa. *Bushveld Complex; Gabbro and norite*.

ATSO VORMA, Geological Survey of Finland, SF 02150 Espoo, Finland. *Rapakivi texture*.

W. J. WADSWORTH, Department of Geology, University of Manchester, Manchester M13 9PL, England. *Cumulate textures*.

GEORGE P. L. WALKER, Department of Geology and Geophysics, University of Hawaii at Manoa, Honolulu, Hawaii 96822, U.S.A. *Explosive volcanic eruptions—classification; Palagonite*.

RICHARD F. WARD, 22434 Melrose Court, East Detroit, Michigan 48201, U.S.A. *Acid(ic) igneous rocks*.

J. L. WARNER, SN7, Chevron Oil Field Research Co., Box 446, La Habra, California 90633, U.S.A. *Lunar petrology*.

JANET V. WATSON, deceased. *Plutonic rocks*.

D. S. WEEDON, Department of Geology and Applied Geology, University of Glasgow, Glasgow G12 8QQ, Scotland. *Skaergaard intrusion*.

A. J. R. WHITE, Department of Geology, La Trobe University, Bundoora, Victoria 3083, Australia. *Anatexis; Andesite line*.

F. WHYTE, deceased. *Basic igneous rocks; Dolerite; Emplacement—modes; Intrusion*.

HAROLD WILLIAMS, Department of Earth Sciences, Memorial University, St Johns, Newfoundland A1B 3X5, Canada. *Ocean-floor metamorphism*.

L. A. J. WILLIAMS, Environmental Science Division, Institute of Environmental and Biological Sciences, University of Lancaster, Bailrigg, Lancaster LA1 4YQ, England. *Rift valley volcanism*.

D. R. WONES, deceased. *Metasomatism; Pneumatolysis*.

A. R. WOOLLEY, Department of Mineralogy, British Museum (Natural History), Cromwell Road, London SW7 5BD, England. *Rheomorphism*.

A. E. WRIGHT, School of Earth Sciences, University of Birmingham, P.O. Box 363, Birmingham B15 2TT, England. *Amphibolite facies; Anorthosite; Breccia pipe; Diatrema; Eclogite facies; Epidote amphibolite facies; Granulite; Greenschist facies; Metamorphic facies; Metamorphic facies series; Metamorphic grade, metamorphic zone; Metamorphic rock composition—graphical representation; Paired metamorphic belts; Polymetamorphism*.

H. P. ZECK, Geological Institute, Copenhagen University, Øster Voldgade 10, 1350 Copenhagen K, Denmark. *Metamorphic convergence*.

The
ENCYCLOPEDIA
of
IGNEOUS AND
METAMORPHIC PETROLOGY

A

ABBREVIATIONS

Ab, An, Di, En, Fa, Fo, Fs, Hy, Ne, Or, Q, Wo, etc.	Calculated normative minerals (see Petrochemical calculations)
al, alk, fm, si, ti, etc.	Niggli numbers (values) (see Petrochemical calculations)
ACF	$A = Al_2O_3 + Fe_2O_3 - (Na_2O + K_2O)$; $C = CaO$; $F = FeO + MgO + MnO$; values plotted as a triangular diagram
AFM	$A = Na_2O + K_2O (Na + K)$; $F = FeO + Fe_2O_3 (Fe^{2+} + Fe^{3+})$; $M = MgO (Mg)$; values plotted as a triangular diagram
A'KF	$A' = Al_2O_3 + Fe_2O_3 - (Na_2O + K_2O + CaO)$; $K = K_2O$, $F = FeO + MgO + MnO$; values plotted as a triangular diagram
CAB	Calc-alkaline basalt(s)
CIPW	Cross, Iddings, Pearson and Washington (norm) (see Petrochemical calculations)
CMAS	CaO–MgO–Al ₂ O ₃ –SiO ₂ system
CPB	Continental plateau basalt(s)
CPX	Clinopyroxene
DI	Differentiation index (q.v.)
DSDP	Deep Sea Drilling Project
HFSE	High-field-strength elements (e.g., Ti, Zr, Y)
I-type	Granites whose magma source regions consisted mainly of igneous material
IAB	Island-arc basalt(s)
IPB	Intraplate basalt(s)
LIL	Large-ion lithophile elements (e.g., K, Rb, Ba, Sr)
LKT	Low-potassium (K) tholeiite(s)
MORB	Mid-ocean ridge basalt(s)
ODP	Ocean Drilling Project
OFB	Ocean-floor basalt(s)
OIB	Ocean-island basalt(s)
OL	Olivine
OPX	Orthopyroxene
P(kb)	Pressure in kilobars
QAPFM	Quartz–alkali feldspar–plagioclase–feldspathoid–mafic (minerals) system
REE	Rare-earth elements

S-type	Granites whose magma source regions consisted mainly of sedimentary rocks
SHO	Shoshonite(s)
T°C	Temperature in degrees Celsius
WPB	Within-plate basalt(s)

Other abbreviations and symbols are given in McBirney, A. R. (1984) *Igneous Petrology* (San Francisco: Freeman, Cooper).

D. R. BOWES

ACID(IC) IGNEOUS ROCKS

The acidic igneous rocks (commonly referred to as “acid” rocks and rarely as acidites) comprise those rocks, whatever their origin and whatever level of the crust in which they crystallize, that are sufficiently silica-rich to contain modal (actual) or normative (theoretical) quartz in quantities of about 10% or more. (Persilicic is a little-used term equivalent to acidic as applied to igneous rocks). Among the volcanic rocks, rhyolite, obsidian, and acidic tuff are the predominant representatives; among the deep-seated rocks, granite and granodiorite are most common.

The term acid (acidic) (along with intermediate, basic, and ultrabasic) is an unfortunate and archaic term and should be replaced by such acronyms as “felsic” or “sialic.” This nomenclature appears to have originated with Elie de Beaumont in 1847. The Sauerstoffs-quotient (“oxygen quotient” or “acidity ratio”) of Bishof and Roth in the middle of the 19th century provides an insight into the terminology. By this classification, the amount of oxygen occurring in RO₂ and R₂O₃ compounds (such as SiO₂ and Al₂O₃) is divided into the amount of oxygen in RO and R₂O oxides (such as CaO, FeO, MgO, and K₂O). The fact that the silicon may form weak acids, while the other cations are base-formers, is presumably the justification for the misleading use of the chemical terms acid(ic) and basic. This antique chemical classification and its many variations are the forerunners of the widely-used CIPW norm (see **Petrochemical calculations**).

The specific rock types included in the acid

rocks are the coarse-grained and plutonic granite, adamellite, and granodiorite. Hypabyssal (dyke) types include microgranite, etc., quartz-bearing porphyries, aplite, and pegmatite. The fine-grained, volcanic representatives are rhyolite, rhyodacite, and dacite. Some other volcanic rocks whose names are based principally on texture, but which usually have normative quartz, are also included. These are felsite, obsidian, pumice, and welded tuff.

Plutonic rocks

Acid rocks make up a considerable portion of plutonic terranes. Batholiths and stocks of granite and granodiorite crop out over thousands of km² of the shield areas and in the cores of fold mountain ranges. They are usually associated with much smaller bodies of intermediate, basic, and alkaline igneous rocks and with large areas of metamorphic schists and gneisses. The acid plutonic and associated rocks are found also to underlie the sedimentary rocks of the continents. These rocks make up the major fraction of the continental masses and their origin and evolution are related to the origin and evolution of the continents.

Petrographically, the acid plutonic rocks are made up chiefly of K-feldspar, Na-rich plagioclase feldspar, and quartz, in proportions that vary with the actual rock type. The K-feldspar may be orthoclase or microcline or, rarely, both. The feldspars are often perthitic, indicating formation of a Na-K-feldspar at high temperature and subsequent exsolution during slow cooling. The principal ferromagnesian minerals are biotite and hornblende, occurring alone or together. Pyroxenes, such as augite and hypersthene, are only rarely present (see **Charnockite suite**). Common accessory minerals are muscovite, magnetite, ilmenite, pyrite, chalcopyrite, sphene, rutile, cassiterite, zircon, apatite, tourmaline, and fluorite. It is rare for any individual body to contain more than a few of these.

Chemically and petrologically there is a gradation with the alkaline rocks on the one hand and with the intermediate and basic rocks on the other. The transition to alkaline rocks is a function of a decrease in the ratio $\text{Si}/(\text{K} + \text{Na})$ that is manifested, mineralogically, by a decrease in the quartz content relative to feldspar. This process ultimately produces the quartz-free monzonites and syenites. A decrease in the ratio $\text{Si}/(\text{Fe} + \text{Mg})$ also accounts for the disappearance of quartz as the acid plutonic rocks grade into diorites. Other chemical variations in this latter case are responsible for the decrease and disappearance of the K-feldspars and the increasing Ca content of the

plagioclase feldspar.

Rock fabric ranges from massive, where there is little or no linear or planar arrangement of mineral grains, to strongly gneissose types that may grade into the associated schists and gneisses. Gradational contacts may range in width from a few meters to several kilometers. Intimately interlayered mixtures are known as migmatite.

Intrusive rocks that are emplaced at shallow depths (ca. 2 km or less) are finer-grained or even glassy, as a result of rapid cooling. They may display distinctly porphyritic texture, e.g., quartz porphyry. These shallow intrusions are gradational with volcanic rocks in character and occurrence.

Volcanic rocks

Acid volcanic rocks are absent, or nearly so, from many volcanic terranes and in only the basalt-andesite-rhyolite association characteristic of orogenic belts do they make up a large fraction of the rocks present. Their occurrence is best seen in light of some consideration of volcanic terranes in general. Historically, many classifications of volcanic regions have been proposed; the following is that of Turner and Verhoogen (1960).

1. Alkaline olivine basalt association, widely developed in continents and in ocean basins.
2. Tholeiitic basalt-quartz dolerite association, characteristically found in a nonorogenic continental environment, but also known to occur in oceanic islands.
3. Leucite basalt-potassic trachybasalt association of nongeosynclinal continental regions.
4. Spilitic-keratophyre association of geosynclines.
5. Basalt-andesite-rhyolite association characteristic of orogenic belts (see **Andesite line; Island arcs-petrology**).

While acid rocks in any great quantity are restricted to the last-named association, if the term is broadened to include any volcanic type having a little normative quartz, then there is a substantial proportion of material in the second category and at least a small proportion in all volcanic associations.

In association (1), acid rocks are insignificant among the huge amounts of basaltic and alkaline lavas; however, some Na-rich rhyolites are known from Samoa. In association (3), a dominantly basic and ultrabasic group, a few occurrences of rhyolites have been described from both the eastern and western African rift valleys in the vicinity of Lake Victoria. In association (4), some rhyolites are known

among the keratophyres of northern New Zealand.

Tholeiitic basalts of association (2) often contain minor normative quartz, which in itself is not a sufficient basis for the use of the term "acid(ic)." However, when extreme cases of differentiation have occurred, such as in the Palisade sill in New Jersey, the resulting rock contains modal quartz and K-feldspar in such proportions as to warrant the use of the term.

Acid volcanic rocks are common and abundant only in the basalt-andesite-rhyolite association with volcanic provinces containing thousands of km³ of intermediate and acid rocks. Some of the more important provinces are the San Juan Mountains of southwestern Colorado, the Absaroka-Yellowstone Plateau in Wyoming, the Cascade Range (which stretches from southern British Columbia to northern California), the Sierra Madre Oriental of Mexico, and much of the circum-Pacific region. These are tectonically active regions in which the rock types range from basalt through andesite to rhyolite. The more acid rocks are overwhelmingly pyroclastic or glassy, being represented by volcanic breccias, tuffs, welded tuffs, ash-flows, ash-falls, obsidian, and pumice. Their distribution and texture suggest that these rocks originated predominantly in violent eruptions of highly viscous fluid or of previously solidified material. This has also been true of historic eruptions in which silica-rich lavas have been subject to more violent eruptions than basalts.

Petrographically, the acid volcanic rocks range from almost entirely natural glass, as in obsidian and pumice, to ones that contain considerable proportions of phenocrysts and clasts in a glassy matrix. In obsidian, the crystalline material present consists largely of extremely small particles of early-formed minerals such as magnetite, which give the glass its dark color. In those rocks in which larger clasts and phenocrysts are present, they are usually quartz, sanidine, biotite, hornblende, augite, sodic plagioclase, tridymite, and orthoclase.

Although these rocks can and do occur as individual, coherent flows, they are far more commonly represented by welded tuffs, tuffs, breccias, lahars, ash-flows, and ash-falls. Because these materials have often been water-worked and redeposited, the geological and stratigraphical relations are usually complex and imperfectly known. Furthermore, rapid weathering of such permeable rocks and the mixing in a single unit of materials from several volcanic sources has made their petrographic and chemical study difficult.

There are, in general, two major theories of

the origin of these acid volcanic rocks. Each relies upon the fact that they occur in regions of active tectonism to provide a mechanism for their origin, and studies of petrogenesis lean heavily on geochemical data, particularly trace-element and isotope data (cf. Dickin, 1981).

1. In areas of major mountain-building activity, granitic magmas are formed by anatexis or metasomatic processes working at depth and some of this liquid finds its way to the surface, producing the acidic volcanic rocks. The associated intermediate and basic volcanic rocks are interpreted as the result of differentiation and contamination of the granitic magma.
2. The basalt-andesite-rhyolite association originates (see **Calc-alkaline rocks**), as do other volcanic associations, with a parental basalt magma. In its ascent through the great thickness of the crust in the mountain belt there is, because of the tectonic mobility and activity, opportunity for the basalt to become contaminated by the silica-rich material in the crust. During this ascent the magma would have to melt and assimilate substantial quantities of siliceous sedimentary and metamorphic rock. Differentiation during crystallization would then give rise to the observed associations.

RICHARD F. WARD

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Cross-references: *Calc-alkaline rocks*; *Charnockite suite*; *Explosive volcanic eruptions—classification*; *Granite*; *Granodiorite*; *Intermediate igneous rocks*;

Petrochemical calculations; Plutonic rocks—classification and nomenclature; Rhyolite; Volcanic rocks.

ALKALINE ROCKS—UNDERSATURATED

Nepheline syenite

Nepheline syenite is a pale-colored, coarse-grained rock composed essentially of alkali feldspar (ca. 70%) and nepheline (ca. 20%) with minor proportions of dark minerals such as sodic pyroxene, sodic hornblende, or biotite; its volcanic equivalent is phonolite. It was first described from Foya (hence the original name, *foyaite*) in S Portugal by Blum (1861). Rosenbusch (1877) proposed the alternative name *nepheline syenite*. An historical review, and a systematic discussion of varieties and related rocks is given by Johannsen (1938). Use of the term *foyaite* has been confused by its adoption by some authors as a name for any nepheline syenite that shows fluidal textures.

Common accessory minerals are sphene (titanite), apatite, and iron ore, but nepheline syenites may carry a large range of uncommon minerals containing elements such as Ti, Zr, Nb, REE, U, Th (Vlasov et al., 1966). The feldspathoid nepheline ($\text{NaAlSi}_3\text{O}_8$) reflects an inadequate silica content in the rock for the complete formation of alkali feldspar ($\text{NaAlSi}_3\text{O}_8$). It may be partly or wholly proxied by other feldspathoids such as sodalite, cancrinite, or haüynite (haüyne). With changing proportion of minerals, nepheline syenites pass into other rock types, some of the more important being shown in Fig. 1. When nepheline exceeds feldspar, the resulting leucocratic rock is called *craigmontite*, and ultimately, *urtite* (feldspar-free). The term *foidite* is used for an igneous rock in which the feldspathoids constitute 60–100% of the light-colored components, e.g., melteigite, ijolite, and urtite (Fig. 1); some authors restrict the term to those igneous rocks in which feldspathoids constitute 90–100% of the light-colored constituents.

The nomenclature of alkaline rocks is both extensive and confused (see **Alkaline rocks—undersaturated: other petrological terms**). A complete list of names and definitions and description of classification schemes are given in Sørensen (1974). The IUGS recommendations relating to names are given by Streckeisen (1976).

Nepheline syenites are largely composed of Na_2O and K_2O and Al_2O_3 and SiO_2 , with only minor or trivial proportions of CaO , MgO and

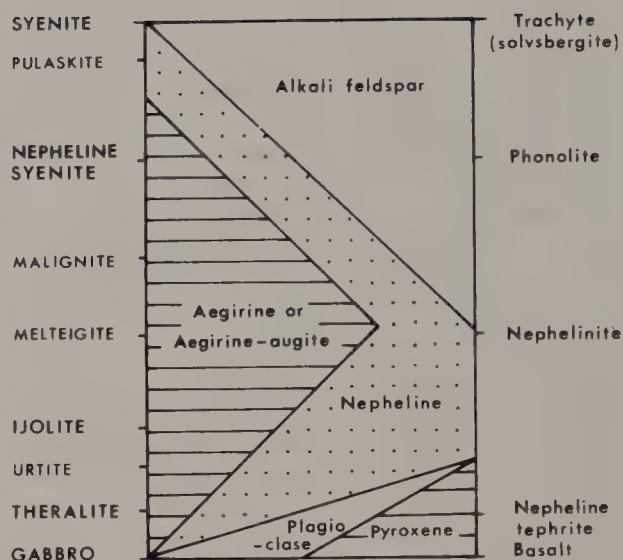


FIGURE 1. Simplified nomenclature and mineral composition of the nepheline syenite and ijolite groups; fine-grained (volcanic) equivalents on right.

Fe_2O_3 – FeO . Johannsen (1938) quotes the following analysis of the type rock from S Portugal (wt.%): SiO_2 55.22, TiO_2 0.59, Al_2O_3 22.59, Fe_2O_3 1.14, FeO 1.17, MgO 0.28, CaO 2.12, Na_2O 8.76, K_2O 5.59, $\text{H}_2\text{O}(+)$ 1.77, $\text{H}_2\text{O}(-)$ 0.39, Cl 0.09, F 0.43 (total 100.27). Average analyses of nepheline syenites and related rocks are given in Nockolds (1954) and Le Maitre (1976).

Nepheline syenites and related rocks fall into two distinct chemical groups determined by the molecular ratio $\text{Na}_2\text{O} + \text{K}_2\text{O}/\text{Al}_2\text{O}_3$. When this ratio is <1 , the rock may be described as aluminous or miaskitic; when molecular alkalies exceed alumina, the rock is peralkaline or agpaite (for full discussion see Sørensen, 1974). This latter condition is noteworthy because it is opposite to that generally found in crustal rocks.

Compared with basalt and granite, nepheline syenite magmas are generated in small amounts within the Earth, but they are an important and characteristic part of the magmatism of the uplifted and rifted portions of the continental shields (see **Rift valley volcanism**).

The compositions of nepheline syenites correspond to low-temperature liquids in the synthetic system $\text{NaAlSi}_3\text{O}_8$ – KAlSi_3O_8 – SiO_2 (“petrogeny’s residua system”—Fig. 2), giving rise to the hypothesis that the natural liquids form at relatively low temperatures, as a result of protracted fractional crystallization or, alternatively, partial melting at shallow depths. There is, however, good evidence that some felsic magmas can, and do, originate in the mantle (Bailey, 1976, 1987, in Fitton and Upton, 1987).

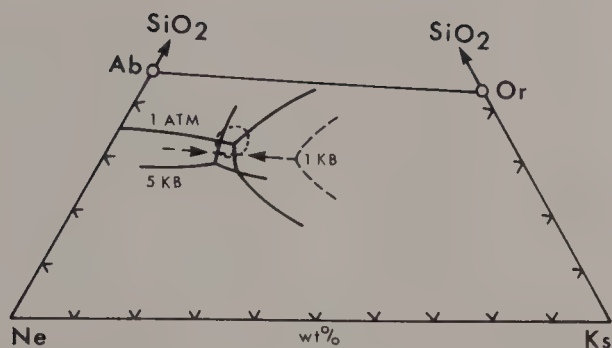


FIGURE 2. Positions of minimum temperature experimental liquids in "petrogeny's residua system" at 1 atm, 1 kb P_{H_2O} and 5 kb P_{H_2O} ; the dashed circle encloses the statistical maximum of analyses of natural nepheline syenites; Ab-Or, alkali feldspars; Ne, nepheline; Ks, kalsilite. (After Bailey, 1976.)

Ijolite

Ijolite is a coarse-grained feldspar-free rock composed essentially of nepheline (50–70%) and pyroxene. When nepheline exceeds 70% the name *urtite* applies and if nepheline is <50% the rock is called *melteigite*. There are no lavas equivalent to urtite and ijolite, but the lava *nephelinite* is equivalent to Ca- and Mg-rich melteigite (see Bailey, 1976). Nephelinite-carbonatite volcanism in E Africa is described by Le Bas (1977, 1987, in Fitton and Upton, 1987).

Minor and accessory minerals are Ca-Fe-Ti garnet, wollastonite, perovskite, sphene, apatite, biotite and Fe-Ti oxides. Textures in the coarse-grained rocks are very variable, and metasomatism of surrounding rocks is common, reflecting a richness in volatiles during crystallization. In nephelinite lavas, olivine and melilite may be present but wollastonite is absent, in contrast to the intrusive rocks. With the incoming of alkali feldspar, ijolite grades into *malignite*, and nephelinite into *phonolitic nephelinite*. In K-rich lavas leucite may take the place of nepheline, giving *leucitite*. As melilite becomes more abundant, nephelinite grades to *melilitite*. Likewise, there are gradations to *hailynitite* and *sodalitite*.

The following is an analysis of ijolite from the type area in Finland (wt%): SiO_2 46.15, TiO_2 0.38, Al_2O_3 15.70, Fe_2O_3 3.55, FeO 3.40, MnO 0.18, MgO 5.52, CaO 14.16, Na_2O 7.24, K_2O 2.16, P_2O_5 0.77, $H_2O(+)$ 0.93 (total 100.59) (Lehijarvi, 1960). Other analyses are in Sørensen (1974).

Ijolitic rocks are highly undersaturated in silica, and in their truly magmatic form (nephelinites) are alkaline ultramafic rocks. Although names such as nephelinite, leucitite, and melilitite are commonly applied, it should be remembered that clinopyroxene is the

dominant mineral, and the rocks are truly pyroxenites. Melilitites represent silicate melts with the lowest recorded SiO_2 contents (lower than peridotite) and another typical member of the ijolite association is *carbonatite* (carbonate magma). These constitute a characteristic part of the magmatism of stable continental plates. Both melilitites and carbonatites show connections with *kimberlites*, which are also characteristically continental. All these silica-poor and alkaline magmas have an origin in the mantle (down to 250 km in the case of kimberlites) involving high activity of mobile elements and volatiles, especially CO_2 (Bailey, 1980, 1984). In many instances the mantle source has been enriched in lithophile elements prior to eruption and mantle fragments may show evidence of metasomatism (reviewed by Bailey, 1982, 1987, in Fitton and Upton, 1987).

Shonkinite

Shonkinite is a dark-colored, coarse-grained rock composed essentially of alkali feldspar (20–30%) and dark minerals, principally augite. Small proportions of olivine, biotite, nepheline and plagioclase may be present, together with accessory iron oxides and apatite, as exemplified by the type rock from the Highwood Mountains, Montana, U.S.A. No equivalent lava has been described. For convenience, shonkinite is shown in Fig. 3 with essexite and theralite, but its real affinities are not clear. Shonkinite has been grouped with *theralite* and *ijolite*, and with *syenite*, but its richness in augite, relatively high K_2O , and regional geologic setting, suggest stronger links with potassic mafic lavas such as those of the Leucite Hills, Wyoming, U.S.A.

Potassic mafic rocks are a highly complex group, having connections with *lamproites*, *lamprophyres*, and *kimberlites*; intrusive varieties such as shonkinite and essexite present especially difficult problems of interpretation. It is impossible at present to assess fully the effects

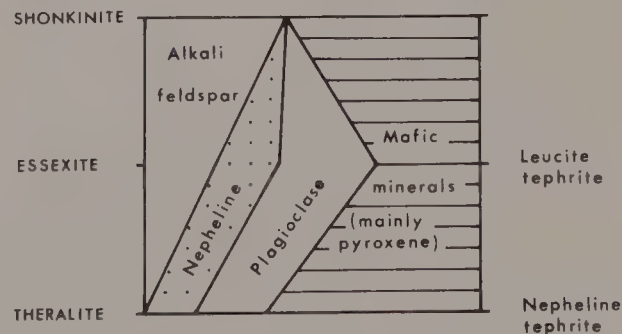


FIGURE 3. Nomenclature and mineral composition of shonkinite, essexite, and theralite, arranged as in Fig. 1. Note that theralite is common to both Figs. 1 and 3.

of alkali mobility in the late stages of crystallization.

Essexite

Essexite is a coarse-grained rock composed of 40–50% dark minerals (chiefly augite) plus plagioclase, alkali feldspar, and nepheline; its volcanic equivalent is leucite tephrite. Most examples contain hornblende and biotite and frequently the nepheline is altered to analcime or other zeolites. Spene, magnetite and apatite are common accessory minerals. Rocks from the type area (Essex County, Massachusetts, U.S.A.) are very variable and have suffered metamorphism and metasomatism. Although the name may not be apt, undersaturated alkali gabbros of this type are reported in small proportions from a wide range of alkaline complexes.

The problems of the origin and relationships of essexite are similar to those of shonkinite, both rocks showing a paradoxical association of apparently lower temperature minerals (alkali feldspar and nepheline) with high-temperature minerals such as titanite.

Theralite and teschenite

The name theralite was originally proposed by H. Rosenbusch for a dark, coarse-grained rock composed principally of augite, plagioclase, and nepheline. Unfortunately, the rock chosen as type example (from Crazy Mountains, Montana, U.S.A.) contains alkali feldspar, not plagioclase, and appears to be intermediate in character between shonkinite and malignite. The mineralogical basis of the original definition has survived, nonetheless, and *thermalite* is now generally used for augite–plagioclase–nepheline rocks. Hornblende, biotite and olivine may be present in theralite (olivine-rich theralite = *kylite*). Magnetite and apatite are the common accessories. In many specimens the nepheline is accompanied or partly replaced by other feldspathoids or zeolites, such as analcime. When analcime completely substitutes for nepheline the rock is called *teschenite*.

Theralite is essentially an undersaturated alkali-rich gabbro and its volcanic equivalent is nepheline tephrite. Intrusions and lavas of this composition are not uncommon in alkali basalt igneous provinces. The origin and relations of theralite are thus part of the more general problem of alkali basalt magmatism.

D. K. BAILEY

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Cross-references: *Carbonatite*; *Feldspathoidal rocks—genesis*; *Haiynite*; *Igneous rocks—tectonic setting*; *Kimberlite*; *Lamproite*; *Lamprophyre*; *Leucitite*; *Melilitite*; *Nephelinite*; *Peralkaline igneous rocks*; *Rift valley volcanism*; *Sodalitite*; *Syenite*.

ALKALINE ROCKS—UNDERSATURATED: OTHER PETROLOGICAL TERMS

In addition to the more commonly used rock names (see **Alkaline rocks—undersaturated**), many local names have been given to varieties of nepheline syenite, ijolite and related rocks, essexite, theralite, and teschenite (Holmes, 1928; Johannsen, 1938; Sørensen, 1974; Bates and Jackson, 1987). Of these the majority are little used, some have been withdrawn and a number are not recommended for use by Bates and Jackson (1987).

The following terms have been used for *nepheline syenites* and related rocks.

Assyntite—an aegirine augite-bearing nepheline syenite with abundant sphene (Assynt, Scotland).

Beloeilite—a sodalite syenite or feldspar-bearing tawite.

Borolanite—a melanite-nepheline syenite with abundant orthoclase and melanite, and subordinate nepheline, biotite, and pyroxene (Loch Borrolan (Borolan), Scotland).

Campanite—a pseudoleucite-bearing effusive equivalent of nepheline syenite; also a variety of leucite tephrite.

Canadite—a nepheline syenite intermediate between albite-nepheline syenite and shonkinite.

Chibinite—a eudialyte-bearing nepheline syenite with a smaller proportion of mafic components than in lujavrite (Khibina, Kola Peninsula, U.S.S.R.).

Congressite—a urtite consisting of nepheline with minor proportions of sodalite, plagioclase, mica, and calcite (Congress Bluff, Craigmont Hill, Ontario, Canada).

Covite—a nepheline syenite with acmite-augite and hornblende (Magnet Cove, Arkansas, U.S.A.).

Craigmontite—a nepheline diorite with nepheline, oligoclase, and muscovite, with minor proportions of calcite, magnetite, corundum, and biotite (Craigmont Hill, Ontario, Canada).

Gooderite—a rock similar to nepheline syenite but with a predominance of albite rather than K-feldspar (Gooderham Township, Ontario, Canada).

Grennaite—a catapleiite-eudialyte-nepheline syenitic rock with a fine-grained matrix.

Husebyite—a plagioclase-bearing nepheline syenite (Huseby, Oslo district, Norway).

Juvite—a leucocratic nepheline syenite with the feldspar exclusively or predominantly orthoclase.

Kakortokite—a banded nepheline syenite with feldspar-nepheline or eudialyte-nepheline and acmite-arfvedsonite bands (Kakortok, SW Greenland).

Kaxtorpite—a pectolite-ekermannite-aegirine-nepheline syenite.

Khibinite—see *Chibinite*.

Ledmorite—a melanite-bearing nepheline syenite with less melanite and more pyroxene than borolanite (Ledmore River, Scotland).

Lujavrite (luijaurite)—a eudialyte-bearing nepheline syenite with igneous lamination (Lujaur-Urt, Kola Peninsula, U.S.S.R.).

Lujavritite—a feldspar-poor lujavrite or feldspar-bearing ijolite.

Malignite—mafic nepheline syenite with >5% nepheline and approximately equal proportions of pyroxene and K-feldspar (Maligne River,

Ontario, Canada).

Midalkalite—an obsolete synonym of nepheline syenite.

Natronshonkinite—a melanocratic nepheline syenite.

Naujaite—a sodalite-rich nepheline syenite with poikilitic texture and microcline with small proportions of albite, analcime, acmite, and Na-amphiboles (Naujakasik, SW Greenland).

Nordsjoite—a variant of juvite without microperthite (Nordsjö (lake), Norway).

Nosykombite—a nepheline covite with barkevikite (Noss-Komba Island, Malagasy).

Orotvite—a nepheline- and cancrinite-bearing diorite with hornblende, biotite, plagioclase, and accessory sphene (titanite), ilmenite, and apatite (Orotva Valley, Detru, Romania).

Pienaarite—a rock showing similarities to malignite but containing anorthoclase as the feldspar and showing exceptional richness in sphene (titanite) (Pienaar Creek, Transvaal, S Africa).

Pulaskite—a leucocratic alkali syenite mainly composed of orthoclase, Na-pyroxene, arfvedsonite and nepheline, although the nepheline proportion may be small; also used for quartz-bearing syenite (Pulaski County, Arkansas, U.S.A.).

Raglanite—a nepheline diorite with oligoclase, nepheline and corundum, and minor proportions of mica, calcite, magnetite, and apatite (Raglan Township, Ontario, Canada).

Rischorrite—a nepheline syenite with nepheline crystals poikilitically enclosed in microcline perthite.

Rutterite—a nepheline-bearing syenite composed mainly of microcline, albite and microperthite; the proportion of nepheline is small (Rutter, Ontario, Canada).

Särnaite—a feldspathoid-bearing syenite with cancrinite and acmite (Särna, Sweden).

Synnyrite—a pseudoleucite-bearing nepheline syenite (Synnyr, Baikal region, U.S.S.R.).

The following are terms used for rocks, particularly *foidites*, related to *ijolite* and its associated rock types.

Aiounite—a melteigite containing titanaugite.

Algarvite—a variant of melteigite with an increased proportion of biotite and a decreased proportion of nepheline.

Arkite—a porphyritic foidite with leucite (or pseudoleucite), nepheline and aegirine augite; similar to fergusonite (see **Leucitite**) but containing melanite (Magnet Cove, Arkansas, U.S.A.).

Fasinite—a melteigite containing titanaugite that is chemically equivalent to berondrite.

Feldspathoidite—a rock group containing the most feldspathoid-rich of the foidites.

Mestigmerite—a mesocratic hornblende-augite ijolite, or a pyroxene malignite (Mestigmer, Morocco).

Niligongite—a plutonic foidite intermediate in composition between ijolite and fergusonite; nepheline and leucite are present in approximately equal proportions; mafic minerals constitute 30–60% (Niligongo, Zaire).

Talzastite—a titanaugite-bearing ijolite (Talzast, Morocco).

Tasmanite—an intrusive rock with a composition similar to that of ijolite, but containing zeolites (natrolite, thomsonite, philipsite, and hydragillite) in place of nepheline (Tasmania, Australia).

Tawite—a plutonic rock similar in composition to ijolite, with 30–60% mafic minerals, but with sodalite in place of nepheline (Tawojak, Kola Peninsula, U.S.S.R.).

The following are terms used for *essexites* and related rocks.

Bjerezite (*bereshite*)—a microleucoessexite rich in nepheline and also containing analcime, titanaugite and aegirine (Bjerez, U.S.S.R.).

Essexite gabbro—a term sometimes used for essexites with $>An_{50}$ in the plagioclase.

Rongstockite—a rock resembling essexite but finer-grained and with less nepheline and sodic rather than calcic plagioclase; also present are orthoclase, cancrinite, augite, mica, hornblende, magnetite, sphene, and apatite (Rongstock (Roztoky), Czechoslovakia).

Sommaite—an essexite containing leucite instead of nepheline and sanidine instead of orthoclase (Monte Somma, Italy).

The following are terms used for *theralites* and related rocks.

Bekinkinite—an igneous rock with little or no feldspar but referred to as a variety of theralite; barkevikite, nepheline, and olivine are major constituents together with Na-amphibole, biotite and analcime (Bekinkina, Malagasy).

Berondrite—a theralite with hornblende and titanaugite as mafic components (River Berondra, Malagasy).

Luscladite—an olivine theralite or essexite characterized by the absence of hornblende; biotite is sometimes present; melanocratic varieties are kyllites (Ravin de Lusclade, Monte Dore, France).

Mafraite—a hypabyssal rock similar to berondrite in chemical composition, containing amphibole, pyroxene, and labradorite (Mafra, Portugal).

Rouvillite—a leucocratic theralite with labradorite (or bytownite) and nepheline as main

constituents; titanaugite, hornblende, pyrite, and apatite are also present (Rouville County, Quebec, Canada).

The following are terms used for *teschenites* and related rocks.

Bogusite—an intrusive rock of the same general composition as teschenite, but lighter in colour (Boguschowitz, Czechoslovakia).

Crinanite—an olivine-analcime dolerite (teschenite) with ophitic texture (Loch Crinan, Scotland).

Cuyamite—a teschenite containing labradorite, analcime, haüyne, hornblende, augite, and magnetite (Cuyamas Valley, California, U.S.A.).

Ijussite—a rock compositionally between teschenite and pyroxenite and composed of abundant titanaugite and barkevikite with smaller proportions of bytownite, anorthoclase, and analcime (Ijuss River, Siberia, U.S.S.R.).

D. R. BOWES

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Cross-reference: *Alkaline rocks—undersaturated*.

ALPINE MAFIC MAGMA STEM

The term *alpine mafic magma stem* was proposed for a distinctive petrological association of alpine-type peridotite and gabbro with Na-rich quartz diorite, trondhjemite, and albite granite, as exemplified by rocks exposed in Canyon Mountain (Thayer, 1963) and near Sparta (Gilluly, 1933) in eastern Oregon. Peridotite and gabbro, which were emplaced together as a solid tectonic block, constitute about 90% of the rocks; the highly sodic siliceous rocks, rich in volatiles and accompanied by extensive albitization, were intruded very soon after tectonic emplacement of the mafic rocks. The postulated magma series is restricted to eugeosynclinal or alpine mountain belts, hence the name; it comprises the plutonic rocks of the ophiolite assemblage. The prototype rocks in eastern Oregon were intruded during Permian to Triassic time, considerably before the Idaho batholith of early Cretaceous

age. The rocks of this magma series in general precede intrusion of large dioritic to granitic batholiths along eugeosynclinal belts.

T. P. THAYER

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Cross-references: *Gabbro and norite*; *Mid-ocean ridge and ocean-floor petrology*; *Ophiolite*; *Peridotite*; *Serpentinite*; *Steinmann Trinity*; *Ultramafic rocks*.

ALTERATION, REPLACEMENT

Processes

Albitization—introduction of, or replacement by, albite, usually replacing a more calcic plagioclase.

Amphibolization—introduction of, or replacement by, amphibole minerals.

Analcimization—replacement of feldspars or feldspathoids by analcime (analcite) usually in igneous rocks during late- and postmagmatic stages.

Autometamorphism—alteration of an igneous rock by the action of its own volatile constituents (e.g., serpentinization of peridotite, saussuritization and uralitization of gabbro); the process is deuteric rather than metamorphic.

Autometasomatism, autopneumatolysis—(see **Metasomatism; Pneumatolysis**).

Biotitization—introduction of, or replacement by, biotite.

Chloritization—introduction of, or replacement by, chlorite.

Damouritization—the process whereby the aluminous silicates (e.g., feldspars) of a rock are transformed into damourite (a variety of muscovite).

Deuteric alteration (synantexis)—alteration in an igneous rock produced during the later stages, and as a direct consequence of consolidation of the magma of the rock; a primary mineral formed by the reaction of two other minerals, as in the formation of a reaction rim, is referred to as being *synantectic*.

Epidotization—hydrothermal introduction of epidote into rocks or rock alteration in which plagioclase feldspar is albitized, freeing the

anorthite molecule for the formation of epidote (and zoisite).

Fenitization—see **Fenite**.

Garnetization—introduction of, or replacement by, garnet—a process commonly associated with contact metamorphism.

Greisenization—hydrothermal alteration by the action of F-bearing water vapor (see **Greisen**).

Kaolinization—the process whereby feldspars and other aluminosilicate minerals are altered to kaolin, commonly by hydrothermal processes.

Mullitization—formation of mullite from minerals of the sillimanite group by heating; a process of very high-temperature contact metamorphism.

Palagonitization—formation of palagonite by the hydration of basaltic glass (see **Palagonite**).

Propylitization (propylite)—(1) hydrothermal alteration of andesite with the development of carbonates, epidote, quartz, and chlorite; (2) growth of white mica reflecting the metasomatic action of hot brines.

Saussuritization—the process whereby plagioclase is replaced by a fine-grained aggregate of albite, zoisite (epidote), and calcite referred to as *saussurite*; the alteration is especially shown by gabbros and in the development of diabase from dolerite; it is commonly accompanied by uralitization and chloritization.

Scapolitization (diprization)—introduction of, or replacement (particularly of plagioclase) by, scapolite, commonly related to the introduction of Cl.

Sericitization—hydrothermal or metamorphic process involving the introduction of, or replacement by, sericite.

Serpentinization—conversion of anhydrous Fe, Mg silicates to hydrous serpentine minerals (see **Serpentinization**).

Silication—conversion into or replacement by silicates, such as in the formation of skarn minerals in carbonate rocks (see **Skarn**).

Silicification—introduction of, or replacement by, silica, generally resulting in the formation of fine-grained quartz, chalcedony, or opal.

Spilitization—hydrothermal albitization of basalt to form spilite (see **Spilite**).

Steatitization—introduction of, or replacement by, talc or steatite, particularly related to the hydrothermal alteration of ultrabasic rocks.

Tourmalinization—introduction of, or replacement by, tourmaline; *tourmalite*, composed almost entirely of tourmaline and quartz, results from metasomatic and pneumatolytic effects along the margins of igneous intrusions and *schorl rock*, composed mainly of aggregates of needle-like crystals of black tourmaline, represents the tourmalinization of granite (see also *luxullianite* in **Granite**).

Uralitization—pseudomorphous development of actinolitic amphibole (*uralite*) after pyroxene, particularly as the result of late-magmatic processes.

Products

Shimmer-aggregate—a micaceous aggregate replacing altered aluminosilicate minerals (e.g., kyanite, cordierite) in metamorphic rocks.

Viridite—a little-used term for the alteration products (e.g., chlorite, serpentine) occurring in scales and threads in the groundmass of a porphyritic rock.

White trap—a term used in the Midland Valley of Scotland for an altered intrusive igneous rock, usually of basic composition, that has been bleached at the contact with coal or other carbonaceous rock; it is developed where gaseous hydrocarbons and CO₂ resulting from the local breakdown of the sedimentary organic matter invade the igneous body at a late stage during its cooling and convert the ferromagnesian minerals and feldspars into a mixture of carbonates and clay minerals.

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AMPHIBOLITE

Amphibolites are hornblende–plagioclase rocks of metamorphic origin that are widely distributed, especially in Precambrian shield areas and in younger orogenic belts. They commonly occur in association with a variety of garnet-, kyanite-, staurolite-, or sillimanite-bearing schists and gneisses and particularly in Precambrian terrane amongst quartzofeldspathic gneisses. Most amphibolites have been produced by the recrystallization of basic igneous rocks under conditions of medium temperature and pressure, although other origins are possible. Well-documented examples occur in the Adirondacks, U.S.A. (Engel and Engel, 1962), Grampian Highlands, Scotland (Wiseman, 1934), Connemara, Eire (Evans and Leake, 1960; Leake, 1964), Broken Hill, Australia (Binns, 1965), and southern West Greenland (Friend et al., 1981).

Metamorphic facies

In common usage, the term *amphibolite* is applied to metamorphosed basic rocks and other hornblende–andesine rocks in the amphibolite facies. According to Miyashiro (1973), such rocks have probably recrystallized under pressures in excess of 4 kb and in the temperature range 600–750°C. Epidote amphibolites, which have the critical assemblage of

epidote, albite, and blue-green hornblende, were formed at lower temperatures. The incoming of orthopyroxene marks the passage to the granulite facies in which a rock with the assemblage hornblende–plagioclase–orthopyroxene is called *hornblende granulite*. A petrogenetic grid for metamorphosed basic rocks is given by Abbott (1982).

Mineralogy

In addition to hornblende and plagioclase, minor proportions of quartz, garnet, biotite, diopside, and cummingtonite may be present. The hornblende is normally green although it may be brown where the metamorphic grade approaches that of the granulite facies. The plagioclase composition varies with metamorphic grade but it is frequently andesine. The garnet is almandine with 6–10% CaO and its distribution is sporadic even in a single outcrop. Diopside forms readily from reactions involving calcite and quartz ($\pm$ tremolite) and is particularly found in amphibolites derived from sedimentary carbonate rocks.

Texture

Many amphibolites are massive medium-grained rocks with equigranular polygonal grains and some show relict igneous textures. Others are foliated (hornblende–plagioclase schist) or strongly lineated. Striped and banded amphibolites with layers 1 mm–2 cm thick represent the products of metamorphic differentiation (Evans and Leake, 1960; Bowes and Park, 1966). Mineral fabric changes with increasing metamorphism are (1) a general increase in grain size, (2) a tendency to equivalence of grain size, (3) a change from ragged to smooth grain boundaries, and (4) a diminution in the number of inclusions.

Chemical composition

The compositions of most amphibolites are very close to those of basic igneous rocks, from which they differ in individual cases in having higher K₂O, H₂O and Fe₂O₃/Fe₂O₃ + FeO ratios. The metamorphism of metabasites is therefore not isochemical, but the compositional changes are small and do not obliterate most of the original chemistry. Where metasomatism is postulated, its extent is normally unspecified. An average water-free composition is (wt%) SiO₂ 50.3, TiO₂ 1.6, Al₂O₃ 15.7, Fe₂O₃ 3.6, FeO 7.8, MnO 0.2, MgO 7.0, CaO 9.5, Na₂O 2.9, K₂O 1.1, P₂O₅ 0.3. Variations in trace element proportions generally reflect the original nature of the rock.

Petrogenesis

There are four main ways by which amphibolite may be formed.

1. Recrystallization of basic igneous rock, whether intrusive, extrusive lava or basaltic tuff, gives rise to *ortho-amphibolites* (*metabasites*).
2. Recrystallization of carbonate–shale mixtures with changes only in the total volatile content (i.e., by loss of CO_2 and H_2O) produces *para-amphibolites*.
3. Recrystallization of some parent material combined with substantial metasomatic addition and subtraction of nonvolatile material produces, in part, para-amphibolites and, in part, skarns.
4. Intrusion and crystallization of basic magma with high $P_{\text{H}_2\text{O}}$ under pressures and temperatures consistent with the amphibolite facies of metamorphism produces autometamorphic or quench amphibolites.

Much effort has been spent in trying to find reliable criteria to discriminate amphibolites of different origins. Evidence for the recognition of para-amphibolites has included layering, their association with marbles, quartzites, and pelites, their lower values of Cr, Ni, and Ti and their higher Niggli k values. However, no criteria based upon element abundance are generally applicable, because some ortho-amphibolites may be low in Cr, Ni, and Ti and some may have their Niggli k value enhanced by alkali metasomatism. Of greater value in the discrimination of the different types are trends of variation of suites of amphibolites compared with known trends of igneous and sedimentary suites. Leake (1964) has shown that on Niggli mg vs. c plots the igneous trend is at a high angle to the variation trend of pelite–carbonate mixtures; amphibolite suites that follow the igneous trend are interpreted as ortho-amphibolites while those suites that correspond compositionally to pelite–limestone or pelite–dolomite mixtures are interpreted as para-amphibolites (Fig. 1).

On this and other bases, most massive amphibolites are ortho-amphibolites as are many banded examples (Evans and Leake, 1960; Misra, 1971). However, Orville (1969) has provided evidence that only pelite–limestone mixtures with Niggli c values of 16–27 will give rise to amphibole-bearing assemblages and, consequently, that any assemblage with dominant hornblende will give an “igneous” trend that is really a mimic of hornblende variability in terms of Mg/Fe (Fig. 2).

To establish a sedimentary trend it is

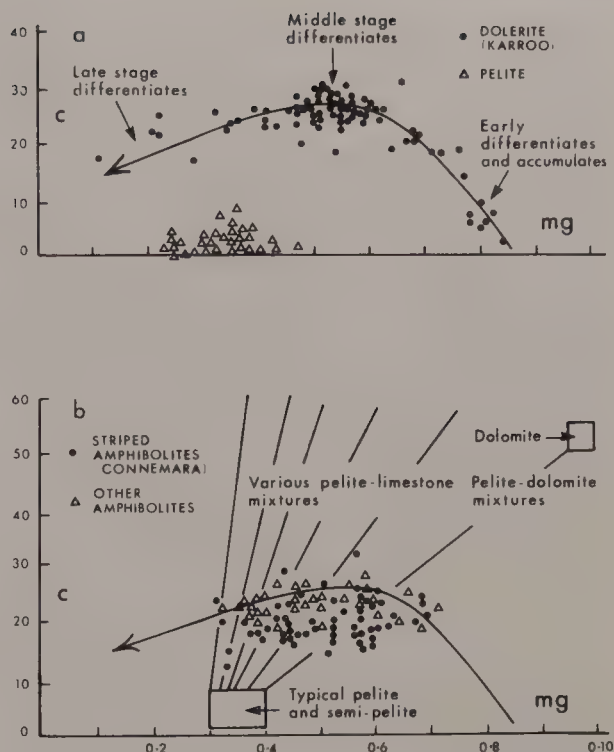


FIGURE 1. Niggli mg vs. c plot ($mg = \text{MgO}/\text{FeO} + \text{MnO} + 2\text{Fe}_2\text{O}_3 + \text{MgO}$); trend line of Karroo dolerites in (a) is transposed onto (b); lines in (b) indicate various limestone–dolomite mixtures mixed with varying amounts of typical pelite and semipelite. (After Leake, 1964.)

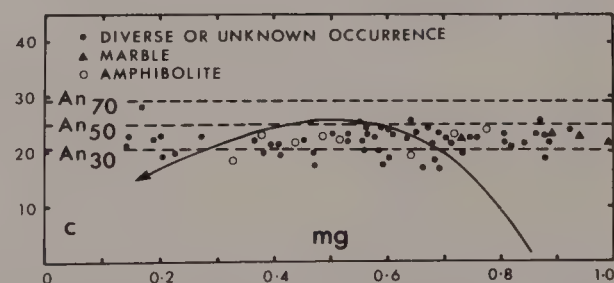


FIGURE 2. Niggli mg vs. c of hornblende analyses from Deer et al. (1963, tables 40, 41, 42, 43); the c values of An_{70} , An_{50} , and An_{30} plagioclase compositions are indicated by dashed lines; the “igneous trend” is shown as a heavy curve and arrow. (After Orville, 1969.)

necessary to examine suites whose compositions extend outside the amphibolite range. Orville proposes that thin-layered amphibolites that exhibit an “igneous” trend may be produced by a bimetasomatic reaction between an original carbonate-rich layer enclosed in a pelite-rich layer. Critical evidence for this process must be sought in the nature of the chemical gradations in the host rock enclosing the amphibolite layers.

D. R. BOWES

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Cross-references: Amphibolite facies; Basic igneous rocks; Epidote amphibolite facies; Granulite facies; Magmatic crystallization in metamorphic environments; Petrochemical calculations; Regional metamorphism; Skarn.

AMPHIBOLITE FACIES

The metamorphic paragenesis of hornblende plus plagioclase in basic rocks is the basis for the original definition of this facies by Eskola (1939). Despite much subsequent discussion, this is still the most satisfactory and widely accepted definition, the lower boundary being marked by the change in plagioclase composition, albite–epidote–actinolite assemblages changing very sharply into oligoclase–andesine–hornblende. The upper boundaries are marked by the appearance of orthopyroxene when

passing into the granulite facies, and by the disappearance of hornblende and the appearance of omphacitic pyroxene into the eclogite facies. The granulite boundary is not normally taken at the disappearance of hornblende as suggested by Lambert (1972) but by the appearance of a distinctive brown aluminous hornblende along with the growth of orthopyroxene. This is not to say that definition of the facies is not difficult, as hornblendes, in these grades, are very variable in composition and definition of the lower temperature limit cannot at present be made in terms of any particular mineral transformations in rocks of basic composition.

Within the stability of this hornblende–plagioclase assemblage there is a wide variety of possible pelitic assemblages (Fig. 1). This has given rise to several suggestions for dividing the amphibolite facies into subfacies. Winkler (1967) divided the stability field of hornblende into three “facies,” viz., almandine amphibolite, cordierite amphibolite and hornblende

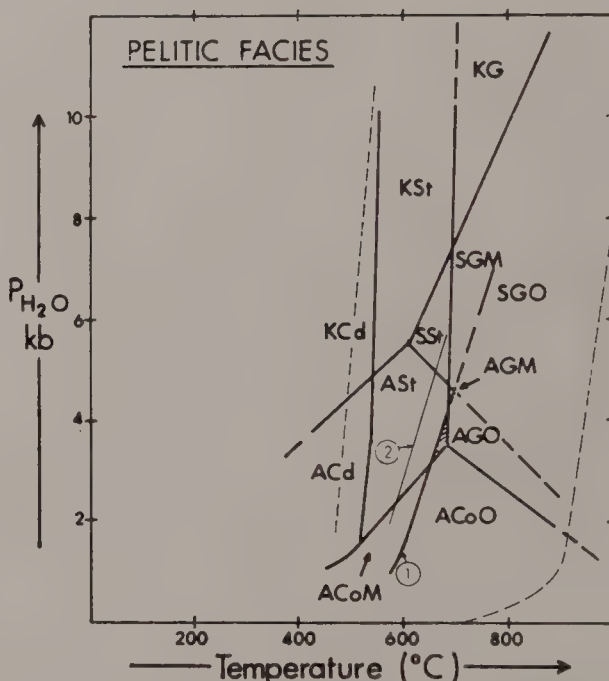


FIGURE 1. Assemblages developed in hydrous, Fe-rich, aluminous rocks (after Lambert, 1972); light dashed lines give stability limits of hornblende; curve 1 is the upper limit of stability of muscovite and quartz; curve 2 is for the reaction staurolite + muscovite + quartz = andalusite or sillimanite + biotite + H₂O. A, andalusite; Cd, chloritoid; Co, cordierite; G, almandine garnet; K, kyanite; M, muscovite; O, K-feldspar (orthoclase); S, sillimanite; St, staurolite. The stability fields are rather indefinite. Porteus (1973) has tried to relate them to the Barrovian and Buchan facies series assemblages in the Dalradian Supergroup in Scotland and has arrived at a slightly different arrangement which is consistent with that particular area.

hornfels. The only mineralogical distinction made between the cordierite amphibolite and hornblende hornfels facies is the presence of wollastonite in calcic rocks in the latter facies. However, it is apparent that workers in areas of Abukumu-type facies series regard cordierite amphibolites as a normal part of the amphibolite facies and would make a distinction from hornblende hornfels largely on occurrence (regional extent) and texture (schistosity as opposed to a hornfels texture).

A completely independent scheme of metamorphic facies published by Russian authors (Dobretsov et al., 1973, 1975) bases the subdivisions on pelitic compositions (see **Metamorphic facies**). There are four facies roughly corresponding to the amphibolite facies. Two higher-pressure facies are defined by the stability of kyanite with the boundary between them defined by the breakdown of muscovite and quartz to give K-feldspar and kyanite. These are termed *kyanite schist* and *kyanite gneiss facies*. The biotite-sillimanite gneiss facies (also called amphibolite facies, although it only occupies a small part of what any other authors have termed amphibolite facies) is the lower-pressure facies defined on the stability of sillimanite (rather than kyanite) in the muscovite + quartz reaction. The lower-temperature and pressure facies is termed the *muscovite-staurolite schist facies* (also the epidote amphibolite facies), being marked by the stability of muscovite, staurolite, andalusite, and cordierite. It is not equivalent to the epidote amphibolite facies defined by any previous authors.

Parageneses in pelitic rocks

It is in pelitic rocks that there is most variation in the amphibolite facies and this gives rise to a number of zones marked by the incoming of the minerals garnet, staurolite, kyanite, or andalusite, sillimanite, and ortho-

clase. The garnet zone is the highest grade of the greenschist or the epidote amphibolite facies, but amphibolite facies assemblages in basic rocks occur towards the top of the garnet zone. Almandine garnet is stable in pelitic rocks throughout the facies, together with biotite. The disappearance of biotite in a reaction resulting in garnet most probably occurs at granulite facies. In Fe-rich pelitic rocks chloritoid is stable for a small temperature range before being replaced by staurolite. At lower pressures, cordierite rather than staurolite is the Fe-Al-bearing phase (Fig. 1). The aluminum silicate polymorphs have a complex relationship. At low pressures andalusite is stable and at higher pressures kyanite is stable. The incoming of kyanite with increasing temperature is apparently at the expense of some staurolite normally and staurolite disappears before the top of the kyanite zone. Kyanite is replaced by sillimanite with increasing pressure. At lower pressures the staurolite accompanies andalusite, but at still lower pressures it is unstable. The first appearance of sillimanite at these lower pressures is controlled by a reaction involving the breakdown of muscovite and the incoming of K-feldspar.

At the upper limit of the amphibolite facies, where orthopyroxene becomes stable, pelitic compositions do not change markedly and compositions in the alumina-rich quarter of the ACF and the A'KF diagrams (Figs. 2c, 3c) are not therefore diagnostic of amphibolite facies.

A model facies system based on pelitic assemblages with quartz and muscovite or K-feldspar, or both, is developed by Thompson and Thompson (1976), although, of the 36 mineral facies they recognize, only four are assigned to the amphibolite facies (of mafic assemblages) rather than the ten major assemblages of Lambert (Fig. 1).

The role of cordierite in rocks of this grade is complex. Chinner (1959) has noted that the cordierite of regional metamorphic rocks is

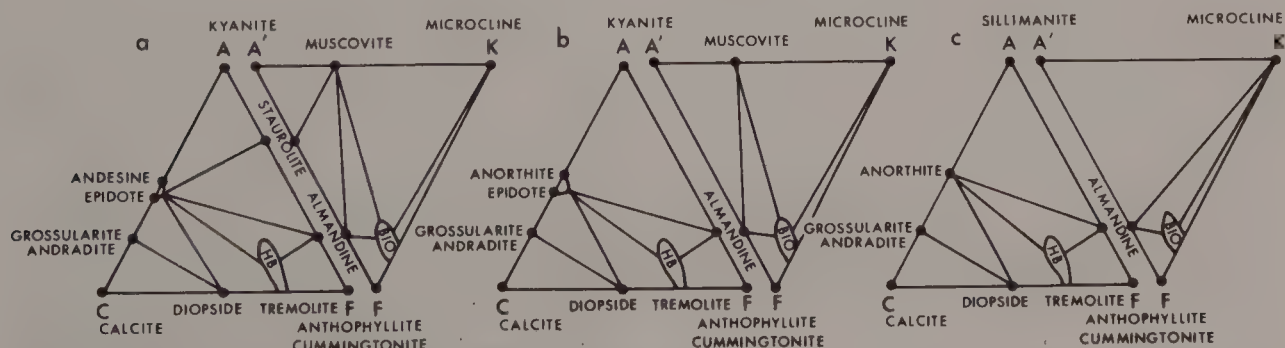


FIGURE 2. (a) Staurolite-bearing low-amphibolite facies, higher pressures; (b) staurolite-free medium-amphibolite facies, higher pressures; (c) sillimanite-orthoclase-bearing high-amphibolite facies, higher pressures. (After Winkler, 1967.)

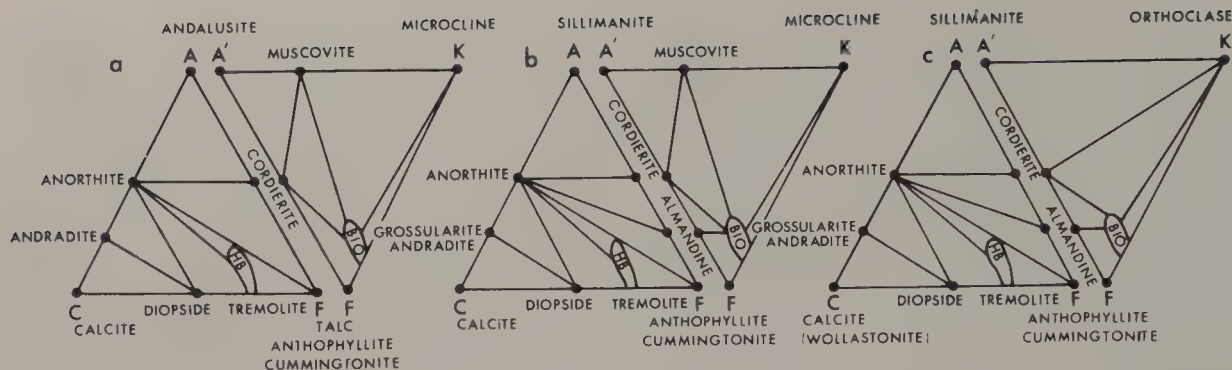


FIGURE 3. (a) Andalusite-bearing low-amphibolite facies, lower pressures; under certain conditions almandine is stable with andalusite without the presence of either staurolite or cordierite (Porteus, 1973); also see Fig. 1; (b) sillimanite-muscovite-bearing medium-amphibolite facies, lower pressures; (c) sillimanite-orthoclase-bearing high-amphibolite facies, lower pressures. (After Winkler, 1967.)

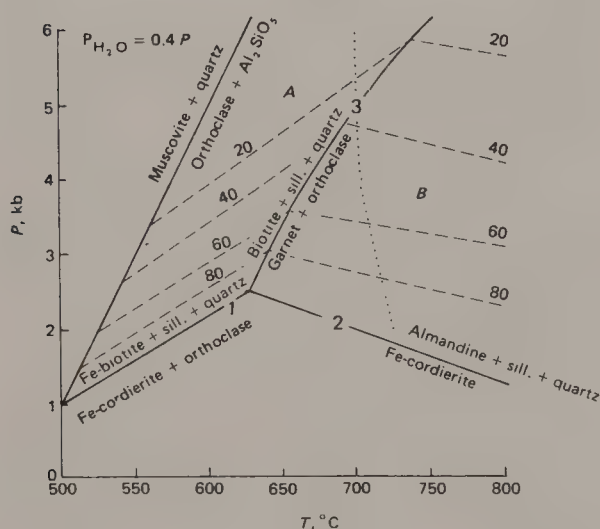


FIGURE 4. Modal molar ratios Fe/(Fe + Mg) in cordierite of cordierite-biotite (field A) and cordierite-garnet (field B) assemblages. (After Holdaway and Lee, 1977.)

more magnesian than that of contact metamorphic rocks. Dallmeyer (1972) shows that for rocks of varying but high grade, the interaction of K_2O , FeO , and MgO is the controlling factor in whether cordierite or garnet or both will crystallize in rocks of sillimanite grade. The complex relationships of cordierite-biotite-garnet and sillimanite are also explored by Holdaway and Lee (1977), who show that the stable cordierite Fe/(Fe + Mg) ratio decreases markedly with pressure (Fig. 4). Thus, it is difficult to predict what cordierite-bearing assemblages would be developed at a particular grade and it is inadvisable to use cordierite as an indicator of major facies divisions.

Calcic assemblages

Both calcite and dolomite are stable throughout the amphibolite facies in conditions of high

partial pressure of CO_2 . At these elevated temperatures and pressures, calcite will take up to 10% $MgCO_3$ in its lattice (Goldsmith 1959), which always exsolves to give intergrowths of dolomite and calcite. In the presence of slight impurities, quartz, or clay minerals, dolomite will generally react to give magnesian silicates, whereas calcite tends to be more stable. The minerals stable at amphibolite facies are forsterite, tremolite, diopside, phlogopite, grossular, and, at lower grades, epidote. The first two are only found in silica undersaturated rocks.

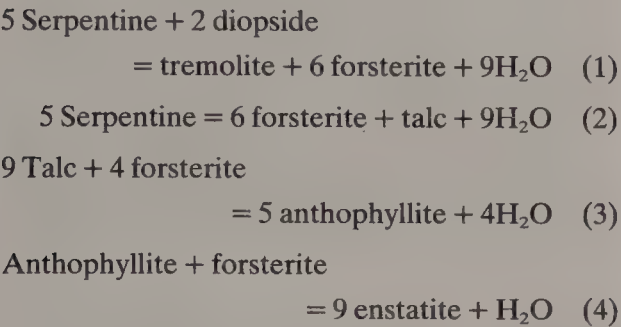
Basic assemblages

As befits the type rock of this facies, basic assemblages do not vary much through this facies. Amphibole is stable throughout as a blue-green or dark green hornblende. Frequently retrogression of higher-grade rocks to amphibolite facies results in sieved green hornblende, or much finer fibrous hornblende aggregates. Generally, it is the most abundant mineral and basic rocks may even consist entirely of common hornblende (hornblendites). Plagioclase makes a marked jump in composition at the bottom of the facies from albite, in greenschist and epidote amphibolite facies, to oligoclase-andesine of composition about An_{28-30} . This jump in composition corresponds to the position of the peristerite solvus (Brown, 1962). With increasing grade there is an increase in the anorthite content of the plagioclase and there is another jump in the composition, in basic and impure calcic assemblages, at about An_{70} (Soper and Brown, 1971). This corresponds to the sillimanite zone of the pelitic assemblages.

Not all basic rocks contain almandine and many contain epidote and, at higher grades, diopside is stable long before the pyroxene granulite facies, which is delimited by stability of orthopyroxene.

Ultramafic assemblages

These are fairly sensitive to temperature changes in magnesian rocks. The following reactions probably all take place within the amphibolite facies, although the major part of the facies lies within the stability field between reactions 2 and 3 (Evans and Tromsdorff, 1970), although some of the reactions (notably reaction 2) are very dependent on XCO₂ (Johannes, 1969):



The lower limit thus more or less corresponds to the stability of tremolite and the upper limit that of enstatite. The presence of anthophyllite indicates high grades and of serpentine indicates low grades. Forsterite is stable throughout the amphibolite facies. The mineral parageneses in the Alps corresponding to these zones and the corresponding assemblages for pelitic and carbonate rocks are shown in Fig. 5.

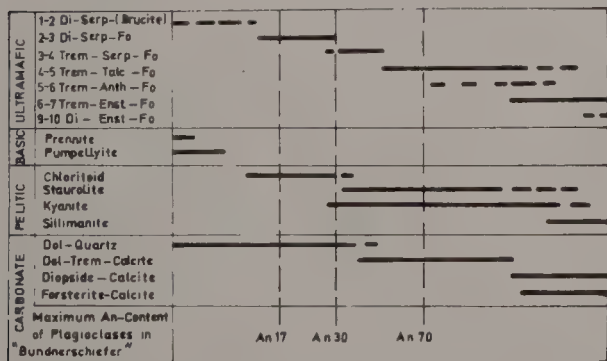


FIGURE 5. Index parageneses in ultramafic rocks of the central Alps and the parageneses of rocks associated with them; the amphibolite facies lower boundary is approximately at the An₃₀ line and the beginning of kyanite and staurolite in pelitic rocks. (After Evans and Tromsdorff, 1970.)

Conditions of formation

Many of the minerals typical of the amphibolite facies may arise from a variety of reactions, so that no specific reactions are applicable to the boundaries of the facies. The reactions involving hornblende at the lower limit follow from a series of reactions involving calcium aluminosilicates, which may be said to be the controlling reactions for the lower grade

facies (i.e., those involving zeolites, lawsonite, pumpellyite, and epidote). When actinolite and hornblende become involved, the range of substitutions possible becomes much larger and natural reactions are difficult to define in the laboratory. Generally, however, on the basis largely of the stability limit of chloritoid, staurolite and, at the upper limit, muscovite and orthopyroxene, the amphibolite facies is probably stable between about 450–500°C and 700–750°C at depths of between 10 and 30 km.

Water pressure seems always to be high. Generally there is no evidence that it was less than load pressure, but the partial pressures of CO₂ and other volatiles may vary considerably. In certain conditions, therefore, scapolite is common, indicating high SO₄ or Cl vapor pressures. At the highest grades, partial melting, and thus the formation of migmatites, is extremely common.

A. E. WRIGHT

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Cross-references: *Anatexis*; *Deserpentinization*; *Eclogite facies*; *Epidote amphibolite facies*; *Granulite facies*; *Metamorphic facies*; *Metamorphic facies series*; *Metamorphic rock composition—graphical representation*; *Migmatite*.

ANALCIMITE

Analcimite is a hypabyssal or intrusive rock composed of analcime and pyroxene (usually titanaugite) with small proportions of feldspathoids and plagioclase, and apatite, sphene and opaque oxides as accessory minerals; olivine may also be present. A monomineralic effusive rock composed entirely of analcime is termed an *analcimolith*. A *blairmorite* is composed of analcime and aegirine augite with minor proportions of nepheline, sanidine and melanite (Blairmore, Alberta, Canada).

D. R. BOWES

ANATEXIS

Anatexis is the process of melting or partial melting of pre-existing solid rocks within the Earth's crust. The term was introduced by J. J. Sederholm in 1907 when discussing one possible origin of migmatites or mixed rocks which consist of both granitic and metamorphic material. The granitic part of the migmatite (leucosome) represents melted material, and the metamorphic part the residual unmelted material (restite). Mainly because of original usage, the term is usually applied when the product of melting is granitic. The term *anatexite* (*anatectite*) is used by some authors, particularly in France, for a rock formed by anatexis.

Anatexis in metamorphic rocks

Explanations other than partial melting have been proposed for the genesis of the leucosomes

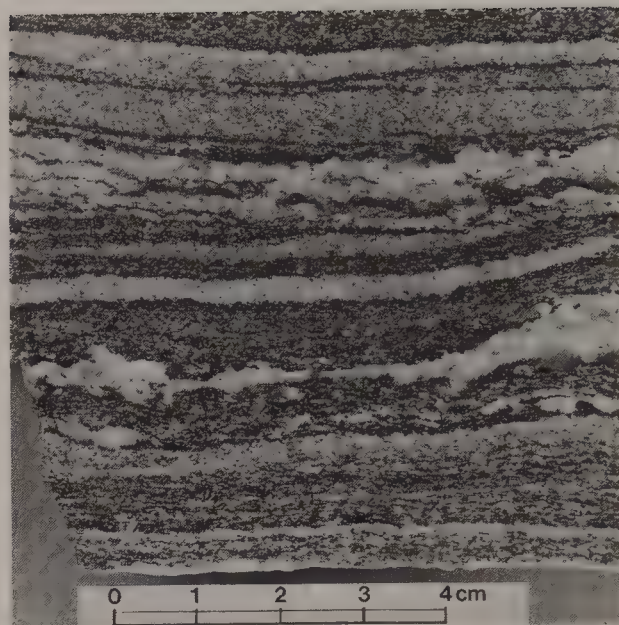


FIGURE 1. Migmatite of the veined gneiss type consisting of alternating bands of granitic composition (leucosomes) and bands rich in biotite; eastern Mt. Lofty Ranges, S Australia.

of migmatites (White, 1966), but most geologists now favor partial melting.

In regional metamorphic terranes, migmatites, mostly of the banded gneiss type (referred to as *lit-par-lit* gneisses, *venites*, or *veined gneisses*—Fig. 1), may occupy many hundreds of km² within the highest-grade parts of these terranes where temperatures were high enough for melting to occur. In the presence of free water, melting must occur in rocks containing quartz + albite + orthoclase at the water-saturated solidus of granite (as low as 650°C at 3.5 kb). Since the melt will be water-saturated, its amount will depend on the amount of free water, as well as the amount of potential granite components in the rock. In a dry system, melt will only be produced at temperatures below the dry solidus and above the saturated solidus when a hydrous phase breaks down to provide the water (vapor-absent melting). For muscovite, this can only occur at pressures above about 3.5 kb, the pressure where the muscovite breakdown curve intersects the granite solidus curve in *P–T* projection (Kerrick, 1972). The melt produced will be water-undersaturated and its amount will depend on the amount of hydrous mineral originally present (Burnham, 1979; Clemens, 1984).

In contact metamorphic rocks surrounding those igneous rocks intruded to high levels in the crust, granitic veins (contact migmatites) are rare. In many contact (thermal) aureoles temperatures are high and commonly above the breakdown of muscovite in the presence of

quartz and albite, but not high enough to be above the saturated granite solidus, which has a shallow negative slope in P - T space at low pressures. Hence, there is normally no melting.

Experimental anatexis

Experimental partial melts have been produced with excess water, at temperatures as low as 660°C and at pressures of only 2 kb, from shales, graywackes and other sedimentary rocks following premetamorphism to assemblages simulating various natural high-grade metamorphic rocks. The compositions of the partial melts range from granitic to granodioritic to trondhjemitic, depending on small variations in starting compositions (Kilinc, 1972).

Some direct experiments involving water-undersaturated systems (either natural or synthetic) have been made (e.g., Brown and Fyfe, 1970) and the present hypotheses for the formation of water-undersaturated granitic melts are based on experimental data as well as theoretical considerations (Burnham, 1979).

Other volatiles may also be involved in anatexis. In water-saturated granite systems, fluorine reduces the temperature of the water-saturated granite solidus appreciably, so that anatexis may occur at lower temperatures and pressures than in pure-water systems (Manning and Pichavant, 1983). Dry and water-undersaturated systems with fluorine and boron are not yet understood. However, Manning and Pichavant (1983) suggest that these elements will probably not promote melting without the presence of water.

Anatexis and granite magmas

Some granites of the large batholiths show evidence of derivation by anatexis of metasedimentary rocks. For example, where cordierite has crystallized from the melt, the magma was water-undersaturated and contained an excess of alumina over that required to crystallize feldspars. These granites may also contain inclusions consisting of biotite, cordierite and aluminosilicate. These are examples of S-type granites (White and Chappell, 1983, and references therein).

Evidence that many of the granites of the large batholiths are not derived from sediments is seen in the presence of hornblende and the abundance of mafic, hornblende-rich inclusions, both of which normally have low initial $^{87}\text{Sr}/^{86}\text{Sr}$. The data are consistent with these magmas being derived by vapor-absent partial melting of preexisting igneous rocks (I-types) near the base of the crust.

Other terms

Anamigmatization—high-temperature, high-pressure remelting of preexisting rock to form magma.

Atexite (atectite)—basic material that is unchanged during anatexis.

Lipotexite (lipotectite)—basic rock material within anatectic magma.

Macula—a local pocket of magma that is formed by the fusion of shale and that acts as a type of magma chamber.

Metatexis (metatexite)—anatexis due to partial melting of rock components with low melting points; the resultant rock is termed a *metatexite (metatectite)*.

Mianthite—dark-colored patches, streaks or enclosures in an anatexite.

Selective fusion—fusion of a portion of a rock with the resultant liquid containing a larger proportion of the more fusible components than the parent.

A. J. R. WHITE

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Cross-references: *Batholith*; *Eutectic melting*; *Migmatite*; *Migmatite—origin of neosome*; *Water in rock melts*.

ANDESITE AND DACITE

Nomenclature and compositional varieties

The name *andesite* was originally coined by Leopold von Buch in 1826 for porphyritic volcanic rocks of the Andes. The rocks were said to consist mainly of "albite" and hornblende. The name was not generally used, however, until a more refined definition was introduced by J. Roth who, in 1861, applied it to porphyritic Cenozoic rocks consisting chiefly of plagioclase and either hornblende or augite. The term was subsequently broadened to include subalkaline (hypersthene-normative) rocks of any age in which plagioclase has an average composition more sodic than labradorite. Modern usage follows the definition adopted by the IUGS Subcommittee on the Systematics of Igneous Rocks (Streckeisen, 1978), which places less emphasis on the composition of plagioclase than on silica content and color index (volumetric proportion of dark minerals). The latter has been set at 35% or less in the mode or molecular norm (40% in the CIPW norm). Many recent classifications set the limits of SiO_2 at 53 and 63 wt%. Rocks in the range of 53–57% SiO_2 are usually referred to as *basaltic andesites*, so that andesite in the strict sense would have a SiO_2 content of 57–63%. The plutonic equivalent of a typical andesite is tonalite or quartz diorite.

The name *dacite* was proposed by Guido Stache in 1863 for certain quartz-bearing volcanic rocks of Dacia, Romania, in which the plagioclase is oligoclase. As now used, the name is ill-defined but is commonly applied to rocks with SiO_2 between 63 and 68 or 69%, a color index of <20, plagioclase more sodic than labradorite, and at least 10% modal or normative quartz. According to the IUGS system, which equates dacite to granodiorite,

the normative feldspar should include between 10 and 33% "alkali feldspar," but there is no clear consensus on how much normative albite is combined with orthoclase to make this component. The criteria of Irvine and Baragar (1971) surmount this ambiguity by using only plagioclase composition and color index, both of which are calculated from the molecular norm (Fig. 1a). *Dacites* are distinguished from trachytes by their greater SiO_2 content for a given composition of plagioclase; although the latter may have sodic plagioclase, most have <10% normative quartz, and many are olivine-normative.

Although the name dacite is applied to subalkaline rocks of many tectonic environments, they are more common in orogenic and continental settings than in purely oceanic regions. This characteristic is even more pronounced for andesites. Orogenic andesites have been divided into two types, tholeiitic and calc-alkaline, on the basis of relative concentrations of Fe and Al_2O_3 . The former are more Fe-rich and have <16% Al_2O_3 . Two commonly used criteria for distinguishing calc-alkaline and tholeiitic rocks are shown in Fig. 1b,c. The name *icelandite* has been used for tholeiitic andesites found in nonorogenic settings, mainly to distinguish them from the rocks of active continental margins and island arcs. No strict compositional distinction has been agreed on, but icelandites tend to be somewhat richer in Fe and poorer in Al_2O_3 than other tholeiitic andesites. TiO_2 is normally <1.3% in tholeiitic andesites and >1.3% in icelandites.

Andesites and dacites have been further classified according to their K_2O contents (Fig. 2). Andesites with <ca. 1% K_2O are often referred to as low-K andesite; they are most common in island arcs, whereas high-K andesites, i.e., those with >ca. 2% K_2O , are confined almost entirely to the interior regions

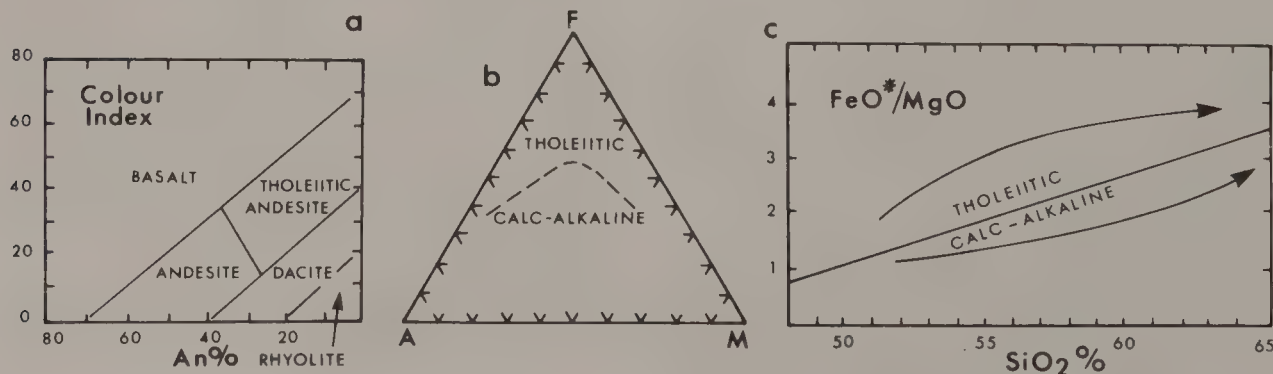


FIGURE 1. (a) Classification of andesites, dacites, and related rocks according to the system of Irvine and Baragar (1971). (b),(c) Divisions between tholeiitic and calc-alkaline rocks according to the criteria of (b) Irvine and Baragar (1971) and (c) Miyashiro (1974); curved arrows illustrate trends of the tholeiitic and calc-alkaline series of the Cascade Range of the northwestern U.S.A. FeO^* , total Fe as FeO .

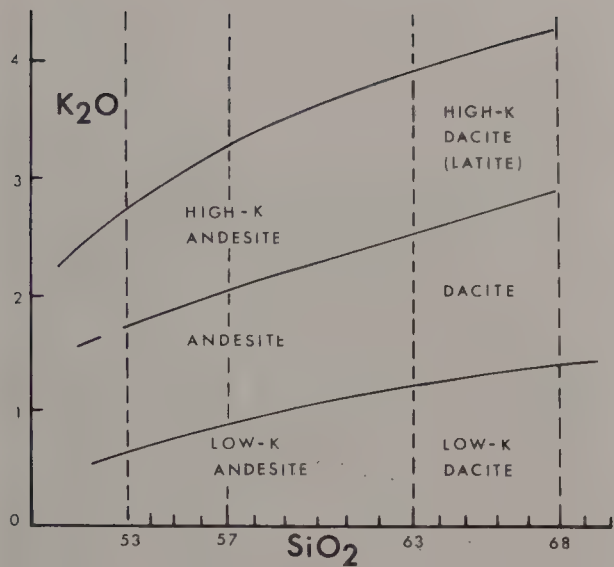


FIGURE 2. Classification of andesites, dacites, and related rocks according to their K₂O and SiO₂ contents; the broken line at 53% SiO₂ separates basaltic andesites from andesites proper. (Modified after Peccerillo and Taylor, 1978.)

of continents. The latter have sometimes been referred to as “shoshonitic” but that term is more strictly reserved for a separate series of very K-rich rocks, most of which have modal K-feldspar. Compositional features of these varieties of andesite and dacite are given in Tables 1 and 2.

Petrographic characteristics

Orogenic andesites are almost universally porphyritic, and the same is true to a lesser degree of dacites. In both, plagioclase, subcalcic augite, and hypersthene are the most common phenocrysts. Olivine may be present as phenocrysts but seldom in abundance and never as part of an equilibrium assemblage of the groundmass. The cores of plagioclase phenocrysts are more anorthitic than the feldspar of the groundmass or norm, and zoning may take a variety of forms, even in the same rock. Many of the calcic andesites (*bandaites*) found in island arcs contain phenocrysts of bytownite and

TABLE 1. Average chemical and normative compositions of the principal types of andesites; major elements recalculated to 100% H₂O free

	1	2	3	4
	Low-K calcic andesite	Calc-alkaline andesite	High-K andesite	Tholeiitic andesite
SiO ₂	58.67	59.00	59.75	59.74
TiO ₂	0.63	0.91	1.05	1.12
Al ₂ O ₃	16.80	17.45	16.86	14.80
Fe ₂ O ₃	3.00	2.64	3.81	1.94
FeO	5.35	4.07	2.78	8.97
MnO	0.15	0.12	0.10	0.20
MgO	3.57	3.41	2.86	2.46
CaO	8.31	6.81	5.49	6.00
Na ₂ O	2.78	3.94	3.91	3.37
K ₂ O	0.62	1.42	2.99	1.17
P ₂ O ₅	0.13	0.23	0.39	0.21
Molecular norms				
Ap	0.28	0.48	0.82	0.45
Il	0.89	1.27	1.47	1.60
Mt	3.18	2.77	4.00	2.08
Or	3.72	8.41	17.73	7.10
Ab	25.35	35.45	35.24	31.02
An	32.02	25.79	19.70	22.29
Di	7.19	5.25	4.07	5.44
Hy	12.06	10.20	6.24	15.78
Q	15.32	10.39	10.72	14.24
Trace elements				
Ba	176	531	1200	225
Zr	61	140	250	140
Rb	8	34	73	30
Sr	287	676	961	150
Co	22	22	22	24
Cr	28	57	83	24
Ni	11	27	68	1

Source: 1,2,3 from Ewart, in Thorpe (1982); 4 from Baker (1978).

TABLE 2. Average chemical and normative compositions of dacites

	1 Low-K dacite	2 Calc-alkaline dacite	3 High-K dacite
SiO ₂	65.79	65.44	66.21
TiO ₂	0.69	0.61	0.63
Al ₂ O ₃	15.63	16.56	16.34
Fe ₂ O ₃	2.53	1.88	2.93
FeO	3.39	2.51	1.05
MnO	0.15	0.10	0.09
MgO	1.79	1.98	1.17
CaO	5.19	4.53	3.14
Na ₂ O	3.81	4.28	4.11
K ₂ O	0.86	1.93	4.11
P ₂ O ₅	0.17	0.17	0.23
Molecular norms			
Ap	0.36	0.36	0.48
Il	0.97	0.85	0.88
Mt	2.68	1.97	1.34
Hm	—	—	1.16
Or	5.15	11.43	24.36
Ab	34.70	38.52	37.02
An	23.34	20.32	14.05
Di	1.32	0.87	0.06
Hy	7.15	6.93	3.21
Q	24.32	18.75	17.44
Trace elements			
Ba	343	646	1587
Zr	n.d.	179	247
Rb	3	62	113
Sr	174	866	637
Co	16	14	8
Cr	10	40	16
Ni	23	30	19

Source: from Ewart (1979).

even anorthite. Pigeonite is restricted to tholeiitic andesites and icelandites, whereas hornblende is more characteristic of calc-alkaline rocks, particularly dacites. Although biotite is found in some K-rich andesites, it is more common in crystal-rich dacites where it tends to be accompanied by phenocrysts of quartz and sanidine.

Many of the phenocrysts are out of equilibrium with the groundmass or with the composition of the rock as a whole. The groundmass of lavas is typically a pilotaxitic or hyalopilitic intergrowth of andesine or oligoclase laths with pyroxene, magnetite, and glass of low refractive index. The proportion of glass tends to be greater in dacites, many of which are vitrophyric. Dacitic obsidian and pumice are common, but andesitic glass is rare. Cristobalite may be found on the walls of vesicles of both andesitic and dacitic lavas; quartz and tridymite occur interstitially in nearly all holocrystalline dacites and many andesites, but quartz pheno-

crysts are much less common in andesites than in dacites.

Occurrence

Andesite and dacite are the most characteristic though not necessarily the most abundant volcanic rocks of orogenic volcanic provinces, such as the island arcs and active continental margins of the circum-Pacific region, Indonesia, and the Lesser Antilles. They are erupted, often explosively, from chains of composite volcanoes associated with trenches and dipping planes of earthquake foci extending to depths of 100–700 km. They are not confined to such settings, however; many are found in the interior regions of continents far from convergent plate boundaries. In the United States, for example, large andesitic volcanoes are found on and around the Colorado Plateau, and dacitic ignimbrites cover wide areas of the Great Basin. In Mexico and Central America both types of rocks are extensive in the plateau regions far behind the main volcanic chain. In Europe, andesites and dacites have been erupted along the northern margins of the Alpine system, especially in the Carpathian Mountains and northern Turkey.

Owing to their relatively high viscosities, the magmas form thick steep-sided flows, many with platy jointing; others are autobrecciated or blocky. Dacite is often extruded as domes, particularly on the upper flanks of large mature volcanoes. Pyroclastic rocks, including air-fall tephra, ignimbrites, and lahars account for much if not most of their total volumes.

Origin

The close association of many andesites and dacites with subduction suggests that the rocks owe their distinctive compositions to some sort of interaction between mantle-derived magmas and crustal material, either of the subducted oceanic lithosphere or the overlying crust. A direct contribution of felsic continental rocks in the form of recognizable xenoliths has been documented in a few instances but is uncommon, and modern interpretations place less importance on this factor than on more subtle contributions in the form of volatile or low-melting components either derived from the subducted slab or scavenged from the overlying lithosphere.

Although liquids with the compositions of andesites or dacites can be produced directly by melting a variety of source rocks under elevated water pressures, it is doubtful that many magmas of these compositions are primary melts. Most of their compositional features can be explained as the result of differentiation of a

basaltic parent. The major-element compositions of both calc-alkaline and tholeiitic andesites are consistent with fractionation of plagioclase, augite, olivine, and magnetite, but trace elements and experimentally determined phase equilibria indicate that amphibole may be an important phase, particularly in the genesis of calc-alkaline andesites and dacites.

Much effort has been devoted to attempts to identify the nature and proportions of material contributed by subducted oceanic lithosphere or continental crust. The isotopes of Sr, Pb, and Nd, which tend to be more radiogenic in orogenic andesites and dacites, may reflect the effects of oceanic crust and sea water. But some of the components of andesites and dacites, such as K, Ba, Rb, and other lithophile elements that have especially high concentrations in magmas that have risen through thick continental crust, seem to be derived from the lithospheric plate above the subducted slab, because their abundances seem to show a better correlation with the nature of the crust than with the nature of the subducted slab.

Other terms

Ambonite—the name of a group of porphyritic cordierite-bearing hornblende-biotite andesites and dacites (Ambonia, Indonesia).

Auganite—(1) a term suggested for an augite-bearing andesite; (2) an olivine-free basalt.

Bandaite—(1) a dacite containing labradorite or bytownite; (2) a quartz basalt texturally like dacite (Bandai San, Japan).

Beringite—a dark-colored barkevikite-bearing andesite or sodic trachyte containing albite and a smaller proportion of orthoclase (Bering Is., U.S.S.R.).

Boninite—a glassy olivine-bronzite andesite containing little or no feldspar (Bonin Is., Japan) (see **Ultramafic lavas**).

Cumraite—a porphyritic effusive rock composed of phenocrysts of bytownite-anorthite in a groundmass of labradorite, pyroxene and abundant glass but with a chemical composition corresponding to andesite rather than to basalt (Great Cumrae, Scotland).

Dacitoid—an effusive rock having the chemical composition of dacite but lacking modal quartz.

Doreite—an andesitic lava containing approximately equal proportions of K and Na; the effusive equivalent of mangerite.

Felsoandesite—an andesite with a felsitic groundmass.

Geburite-dacite—a holocrystalline variety of dacite occurring as a dyke rock, and characterized by the presence of hypersthene.

Hungarite—andesite containing abundant hornblende.

Inninmoreite—a porphyritic rock similar to cumraite but with the plagioclase phenocrysts being anorthosite-labradorite and plagioclase in the groundmass being sodic (Inninmore, NW Scotland).

Kohalaite—an andesite that contains normative oligoclase; it may or may not contain modal olivine (Kohala Mt., Hawaii).

Leidleite—a glassy variety of dacite or rhyodacite containing microlites of calcic plagioclase and pyroxene in a groundmass with subvolcanitic texture.

Orthoandesite—an andesite containing orthopyroxene.

Sakalavite—a glassy andesite containing plagioclase and augite phenocrysts and quartz xenocrysts in a glassy groundmass.

Sanakite—a glassy andesite composed of bronzite, augite, magnetite, and some large plagioclase and garnet crystals—a garnetiferous boninite (San-uka, Japan).

Santorinite—(1) a hypersthene andesite containing plagioclase crystals with labradorite cores and sodic rims and a groundmass with microlites of sodic oligoclase (Santorini, Greece); (2) a light-colored effusive rock with ca. 60–65% SiO₂ and calcic plagioclase (labradorite-anorthite) as the only feldspar.

Sanukite—a glassy andesite with orthopyroxene and andesite crystals (San-uka, Japan).

Shastaites, *shastalite*—obsolete terms originally applied to dacite and unaltered glassy andesite, respectively (Mt. Shasta, California, U.S.A.).

Suldenite—a hornblende andesite.

Tomazite—a greenstone-like, metamorphically altered hornblende-biotite andesite (Timok Valley, Siberia, U.S.S.R.).

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Cross-references: *Andesite line*; *Calc-alkaline rocks*; *Circum-Pacific plutonic and volcanic activity*; *Diorite and tonalite*; *Explosive volcanic eruptions—classification*; *Granodiorite*; *Igneous rocks—classification and nomenclature*; *Igneous rocks—tectonic setting*; *Intermediate igneous rocks*; *Island arcs—petrology*; *Ultramafic lavas*; *Volcanic rocks*.

ANDESITE LINE

Andesite with distinctive calc-alkaline chemical composition and mineralogy (Table 1; Taylor and White, 1965) is the characteristic lava of the Quaternary volcanoes of the circum-Pacific island arcs. This was recognized at the turn of the century when andesites and associated rocks

TABLE 1. Chemical composition of an andesite typical of those along the andesite line and of oceanic rocks sometimes called andesites

	1	2	3	4
SiO ₂	57.4	48.67	51.99	59.64
TiO ₂	0.65	2.28	3.02	1.74
Al ₂ O ₃	17.7	15.91	16.30	14.61
Fe ₂ O ₃	3.20	3.24	2.75	4.28
FeO	3.30	7.13	7.44	3.88
MnO	n.d.	0.15	0.11	0.14
MgO	3.45	7.72	3.19	2.10
CaO	7.40	8.02	6.67	4.68
Na ₂ O	3.40	4.26	5.64	4.24
K ₂ O	1.25	1.28	2.13	1.96
P ₂ O ₅	n.d.	0.62	1.25	0.47
H ₂ O	1.60	1.06	0.36	1.96
TOTAL	99.35	100.34	100.85	99.70

- 1. Andesite; Viti Levu, Fiji; Taylor et al. (1969).
- 2. Hawaiite (andesine andesite); Kohala volcano, Hawaii; Macdonald and Katsura (1964)—contrast the low SiO₂ and high Fe/Mg ratio, with that of andesite.
- 3. Mugearite (oligoclase andesite); Kohala volcano, Hawaii; Macdonald (1949)—contrast the low SiO₂, high alkalis, and high Fe/Mg with typical andesite.
- 4. Icelandite (tholeiitic andesite); Jervis Island, Galápagos; McBirney and Aoki (1966)—notice that CaO and Al₂O₃ are lower and there is a high Fe/Mg compared with typical andesite even though SiO₂ is similar.



FIGURE 1. The andesite line of Marshall (1912)—M and Taylor et al. (1969)—T.

were referred to as the “Pacific Suite.” In contrast, products of intra-Pacific volcanoes are either alkaline or tholeiitic or a mixture of both. Along the eastern side of the Pacific, the boundary between these petrographic provinces is the continental margin of the Americas, but in the western and southwestern Pacific region, a line may be drawn between islands containing contrasting petrographic rock types. This boundary line is known as the *andesite line* (Fig. 1).

Historical development

Marshall (1912) was the first to draw a boundary, corresponding to the andesite line, on the map of the SW Pacific. His line was essentially drawn on the basis of known bathymetric data but he maintained that structural, petrological and biological characteristics lent support to his contention that this was the “true boundary of the Pacific basin.” He stated that “. . . in all the island groups lying to the east of this line only basic and alkaline rocks are known.”

In 1944 Bryan referred to this as the “Marshall line,” although later it was sometimes known as “Marshall’s andesitic line.” The term andesite or andesitic line is attributed to an obscure publication by Born in 1933 (see Bryan, 1944; Gutenberg and Richter, 1954). Except for minor differences in detail, Born’s line based on petrographic criteria, and a more recent line (Taylor et al., 1969) based on geochemical data and constructed independently of earlier workers, are both very close to Marshall’s line.

It should be noted that when Marshall first published on this subject in 1912, intermediate rocks of the alkaline associations were not referred to as oligoclase or andesine andesites (now mugearites and hawaiites) nor had tholeiitic rocks, including tholeiitic "andesites" (now icelandites), been recognized amongst rocks of the oceanic islands. These alkaline and tholeiitic intermediate volcanic rocks have sometimes been confused with calc-alkaline andesites and this, in turn, led to confusion about the nature of the andesite line. Although each of these rock types consists essentially of various pyroxenes and plagioclase, the calc-alkaline andesites characteristically have some groundmass hypersthene, whereas hawaiites and mugearites only contain clinopyroxenes and the pyroxenes of icelandites include pigeonite. There are also differences in chemical composition between all these rocks (Table 1).

Significance

Although Marshall emphasized his conclusion that the line he drew represented the margin of the Pacific basin, he also said that it "may in the past have formed the extension of the Australian continent." Later workers up until about 1960 (e.g., Gutenberg and Richter, 1954) followed this suggestion and considered that the andesite line in the western Pacific was along the continental margin and that the ocean west of the line was floored by foundered continental crust.

In the period 1960–70 it was shown that at least some of the enclosed seas between the andesite line and the continental land masses have oceanic crust (e.g., Philippine Sea and Tasman Sea). Karig (1970) proposed that these marginal basins result from extension and formation of new oceanic crust as the frontal arc–trench system, and hence the andesite line, migrates away from the continent with time.

According to the plate-tectonic model, the andesite line corresponds to the convergent western margin of the Pacific plate (Fig. 2), the upper part of which is composed of oceanic crust. At convergent plate margins involving plates with oceanic crust, the approximate planar arrangement of earthquake foci dipping beneath the volcanic arc (Benioff zone) is thought to be caused by a downgoing slab of lithosphere. The calc-alkaline andesites associated with these margins are considered to have been produced by partial melting of hydrated oceanic crust on the upper part of the lithospheric slab. If andesites are produced by this mechanism, old continental crust is not necessary for their formation. This means that the andesite line along the western Pacific is not

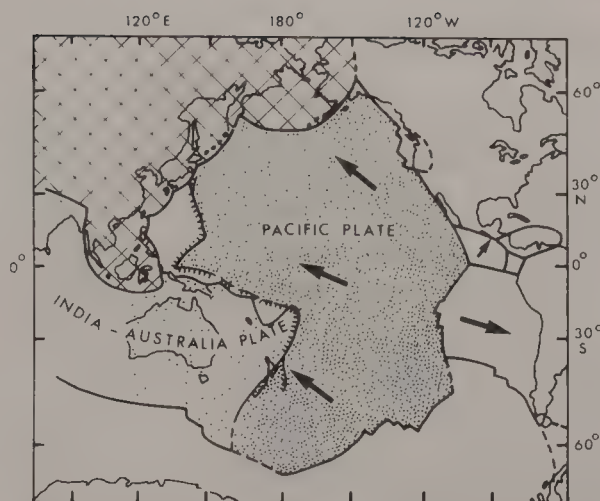


FIGURE 2. Relationship between the boundary of the Pacific plate (plate boundaries in solid lines) to Gutenberg and Richter's (1954) andesite line (hachured line); arrows indicate movement directions of plates.

the edge of the Australian continent, nor is it the boundary of the Pacific basin as originally suggested by Marshall in 1912, but the boundary of the Pacific plate along which andesite magmas are produced.

A. J. R. WHITE

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Cross-references: *Andesite and dacite; Calc-alkaline rocks; Circum-Pacific plutonic and volcanic activity; Island arcs—petrology; Pacific and Atlantic suites.*

ANORTHOSITE

Anorthosite is a coarse-grained rock composed essentially of plagioclase feldspar in the range andesine to anorthite. The term was coined by T. S. Hunt in 1863 to apply to rocks composed of "anorthose" feldspars, an obsolete term for triclinic feldspars. Anorthosite is now universally applied to plagioclase-rich rocks of whatever origin, although no single quantitative mineralogical definition is in general use. Anorthosites are polygenetic and there are many different origins, igneous, metamorphic and metasomatic, and all present problems in petrogenesis. For igneous anorthosites there are three variables to consider in the definition of the rock, viz. (1) plagioclase %, (2) composition of the plagioclase, and (3) quartz %. The minimum % of plagioclase is variously recommended as 85, 90 or 95%, 90% being the preferred usage (Buddington 1939; Streckeisen, 1967). Generally with a fall in plagioclase % the rock grades into gabbroic anorthosite or norite. In differentiates of basic rocks, the plagioclase is labradorite or bytownite, but in anorthosite massifs the composition of the plagioclase commonly falls below An_{50} and may be as low as An_{35} (Buddington, 1939). However, rocks composed of albite or oligoclase are never included within the term anorthosite and they are usually found in quite different geotectonic environments and are of different origin from other plagioclase-rich rocks.

Anorthosites with 5–10% quartz are possible and these are normally the more sodic varieties. They grade, therefore, into quartz-bearing diorites, but as the quartz diorites must have at least 20% quartz (Streckeisen, 1967) there is a distinct difference in quartz content between all anorthosites and quartz diorites.

Three main types of anorthosite occur, of contrasted composition, environment and association: (1) those found in layered basic and ultrabasic igneous complexes, termed the *layered-type*; (2) those found as plutonic intrusions of batholithic aspect in orogenic belts and termed the *massif-type*; and (3) those found in gneissose bands in high-grade metamorphic rocks and termed the *metamorphic-stratiform-type*. The third type may in places be recognized as metamorphosed relics of the first type. In addition, anorthosite blocks have been recovered from the ocean floor in the vicinity of mid-ocean ridges, and from the Moon (see **Lunar petrology**).

Layered-type

Two varieties of layering are recognized in layered intrusions—Skaergaard-type and Willow

Lake-type (Taubeneck and Poldervaart, 1960). There are many examples of intrusions with Skaergaard-type layering but the most famous are the basic-ultrabasic complexes typified by the Bushveld Complex. Very old banded anorthosites have also been widely recognized and these are exemplified by the Fiskenaesset Complex, W Greenland.

Besides the occurrence as distinct bands in well-differentiated bodies, anorthosite also occurs as bands and lenses in complexes that are otherwise poorly differentiated. Indeed, feldspar-rich concentrations of very small size are found in almost every gabbroic complex. Alpine-type intrusions may also contain centimeter-scale layered anorthosites; these are very widespread in both age and location.

Although the exact mechanism of differentiation is not known, it is apparent that the differentiation takes place more or less *in situ* and, because of this, anorthosite is extremely rare in multiple or composite intrusions (Irvine, 1982). An exception is the occurrence of large anorthosite lenses injected as separate plutons of the Nkenza gabbro of the Livingstone Mountains, southern Tanzania (Wright, 1963). This is a series of separate intrusions of olivine gabbro, troctolite, and norite, the first intrusions being a series of anorthosites and pure magnetite rocks. It is chemically unusual also, the plagioclase (An_{52}) being much more sodic than is usual even for anorthosites forming high in layered basic complexes.

Bushveld-type. These occur in large layered basic complexes such as the Bushveld, Stillwater, and Muskox intrusions, in which igneous differentiation has given rise to monomineralic bands (Wager and Brown, 1968). The banding is usually horizontal and such masses are probably the largest single intrusions found on the Earth (see **Bushveld Complex; Intrusion-layered**; see also **Skaergaard intrusion**).

The intrusions are thousands of meters thick and several hundred kilometers in diameter. The anorthosites occur in distinct zones in which they predominate or are very common. They are interbanded with other more or less monomineralic rocks, such as bronzitites, dunites, and chromitites, and with variable gabbros and norites. Sedimentation structures, microscopic textures and the phase petrology provide conclusive evidence of the igneous origin of the rocks and of the crystal fractionation of the magma. However, the exact mechanism of differentiation of the anorthosite has not been solved, the great thickness of individual anorthosite layers—up to 450 m in the Stillwater Complex (Hess, 1960, in Wager and Brown, 1968)—being the most difficult fact

Bushveld Intrusion South Africa			Skaergaard, East Greenland and Stillwater, Montana		
Thickness, ft	Overlying rocks	Irregular masses of melanogranophyre	Thickness, ft	Upper border group	Gabbro and ferrodiorite
~1,900	Upper zone	"Ferrogabbro" or ferrodiorite with melanogranophyre concentrations	2,800	Upper zone	Ferrodiorite or "ferrogabbro" with granophyric concentrations
			2,300	Middle zone	Gabbro
			2,400	Lower zone	Olivine gabbro
			Sk. St. = ? = ?	Gap or overlap uncertain	Base unexposed
			2,100	Upper gabbro zone	Top unexposed Gabbro
11,700	Main zone	Gabbro and anorthosite (poor layering)			
			6,200	Anorthosite zone	Anorthosite
					Gabbro
					Anorthosite
					Gabbro
					Anorthosite
3,400	"Critical series"	Anorthosite, norite, feldspathic pyroxenites, orthopyroxenite (fine layering)	2,200	Lower gabbro zone	Gabbro
	Chromite horizon	Chromite	2,700	Norite zone	Anorthosite, norite, feldspathic pyroxenite fine layering
4,000	Basal series	Orthopyroxenite harzburgite, dunite	1,100	Chromite horizon	Orthopyroxenite
		Peridotites			Chromite
			2,400	Ultramafic zone	Orthopyroxenite, harzburgite, minor dunite
400	Marginal zone	Fine-grained norite			Peridotites
			500	Border zone	Feldspathic orthopyroxenite

FIGURE 1. Thicknesses of the main rock types in the Bushveld Complex compared with those in the Stillwater Complex; the Skaergaard intrusion is regarded as roughly equivalent to the top part of the Bushveld Complex. (From Hyndman, 1972.)

to reconcile with considerations of mineral density and magma composition (Fig. 1). Irvine (1982) suggests that while crystal settling is a possible mechanism, others are flotation accumulation and formation of crystal "clouds" suspended in the magma either by low density contrasts between crystals and melt, by high melt viscosity, or by the presence of barriers in a density-stratified liquid body (see **Double-diffusive convection**).

The composition of the feldspar in these complexes varies with the stage of fractionation. In the Stillwater Complex the composition varies through the anorthosite zone from An_{79} to $An_{75.5}$. There is a much greater difference when more than one anorthosite zone occurs, as in the Bushveld Complex. The anorthosites of the Critical Zone in the lower part of the complex are of bytownite (An_{80-90}), while those in the Main Zone, near the top of the complex are of labradorite (An_{59-65}). There is a clear distinction in the Bushveld Complex between these two types and the anorthite percentage of anorthosites thus gives an indication of the stage of differentiation reached. The Critical Zone anorthosites are associated with bands of pure chromite and are the predominant rock type of this part of the sequence. The associated rocks are norites and bronzitites. The Main Zone

anorthosites are associated with bands of pure magnetite (with again no transitional types) but do not form such an important part of the complex. The associated rocks are bronzite norites and gabbros. Two such types of anorthosite can be recognized in other layered complexes although frequently only one of the two types is developed in any particular complex.

The texture of this type of anorthosite is normally coarse-grained, and usually larger than that of the other rock-types. The large plagioclase crystals have interstitial grains of hypersthene, clinopyroxene or iron ore. In the Stillwater Complex there are also plagioclase grains interstitial to the large euhedral plagioclase. The ferromagnesian constituents may also occur as clusters of grains that give rise, according to the size and shape of these clusters, to spotted, mottled or streaky anorthosites.

Fiskenaesset-type. In early Precambrian time (more than 2,500 Ma ago) there was widespread intrusion of large layered complexes with predominating very calcic anorthosites and chromitites. These are best developed in W Greenland (Fig. 2; Windley, 1967; Windley et al., 1973, see Windley, 1984) but equivalent types probably of similar age are found in many areas of early Precambrian rock (e.g., Limpopo

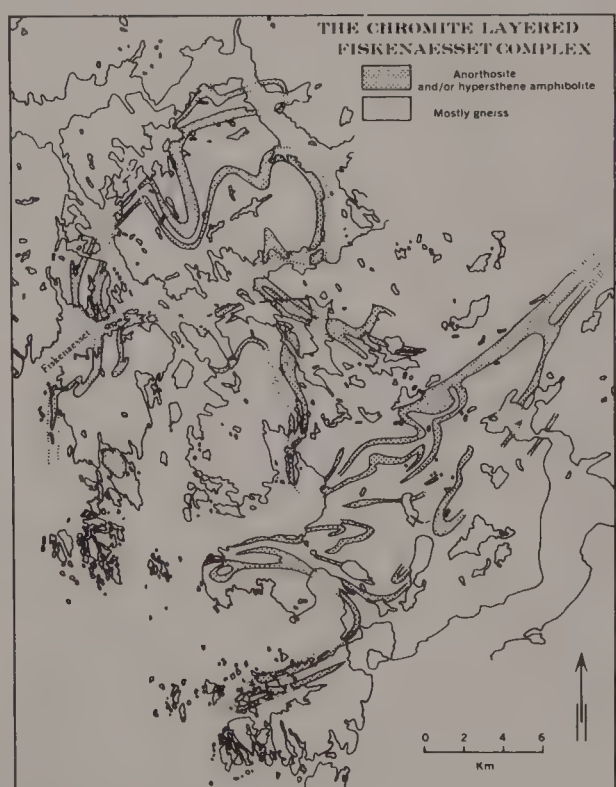


FIGURE 2. Map of the Fiskenaesset Complex, W Greenland, showing the extent of the layered complex (125 km long) and the complex refolding. (From Windley, 1969.)

Belt, South Africa; Madras, India; Western Australia).

In the Fiskenaesset region of W Greenland cryptic variations of mineralogy and cumulate textures are still preserved even after isoclinal folding and granulite facies metamorphism (Windley, 1984 and references therein). The plutons were probably lopoliths of a hydrous high-Al basaltic parentage. The very calcic nature of the plagioclase (An_{80-96}) is probably accounted for by the high water pressure of the magma. It would also lead to the crystallization of primary amphibole, which would remove less Ca and more Na from the melt than clinopyroxene from a corresponding anhydrous magma. The layered sequence is thus highly dissimilar to the Bushveld-type sequence, as well as being very highly deformed and metamorphosed.

Willow Lake-type. Layering in these complexes is a vertical banding parallel to the sides of cross-cutting batholithic intrusions. It is found rarely in gabbroic plutons and gives rise to normal labradorite anorthosite as thin monomineralic layers. In dioritic complexes anorthosites of much more sodic composition may develop.

Ophiolite and ocean-floor anorthosites. Anorthosite layers also occur to a very limited extent in the layered complexes developed at mid-oceanic ridges (Engel and Fisher, 1969) and the crystallization mechanism of such magma chambers and the nature of plagioclase accumulation there has provided one of the best models for anorthosite genesis in layered complexes in general (Lister, 1983). They are, however, extremely rare and only a few examples of anorthosite are known from ophiolite complexes (Coleman, 1977, p. 37).

Massif-type

These occur in large plutonic massifs that may cover many thousands of km^2 (Fig. 3; Isachsen, 1969). The classic pluton of this type is the Adirondack anorthosite (Buddington, 1939) but it has been suggested that there are two massif-type anorthosites, labradorite-type and andesine-type (see Anderson and Morin, in Isachsen, 1969). The Adirondack pluton is of andesine-type. The differences between the two are given in Table 1. Many of the labradorite-type plutons are layered, e.g., Michikamau (Emslie, in Isachsen, 1969) and so may be more comparable to the layered-type. Gabbroic anorthosite (i.e., with plagioclase 75–90% of the rock) is the predominating type in labradorite-type massifs, whereas the andesine-type has a majority of true anorthosite. Other associated rocks are of great importance in the



FIGURE 3. Massif-type anorthosites in Canada. (After Windley, 1984.)

andesine-type, the whole kindred from anorthosite to *mangerite* and *charnockite* (of the charnockite suite) being characteristic. The occurrence of the more acidic varieties of the anorthosite–charnockite kindred is most probably dependent on the prevailing pressure and water vapor pressure (Fig. 4), possibly reflecting partial melting in the envelope rocks (see **Charnockite suite**). They typically occur in granulite facies rocks.

Other associated rocks are usually of minor importance, although a few have small bodies of monomineralic rocks, e.g., pyroxenites and particularly titanomagnetite rocks. The latter are particularly useful as indicators of temperature and pressure of formation (Anderson and Morin, in Isachsen, 1969).

These complexes are apparently confined to Precambrian orogenic belts, and many have metamorphic mineral assemblages. Because of the reaction between ferromagnesian minerals and the feldspar, garnet is then abundant and, gauged solely on feldspar percentage, many of these masses are not anorthosite. Chemically, however, they are anorthositic and the term meta-anorthosite is widely used for this type of rock.

They are characterized by plagioclase compositions between An_{40} and An_{65} , the most common type, the andesine-type, being near the andesine–labradorite boundary. The Adirondack anorthosite is not entirely typical in this respect, varying from An_{38} to A_{49} with the

TABLE 1. Features of two types of massif-type anorthosite

Attribute	Labradorite-type	Andesine-type
Feldspar	Labradorite (An ₆₃₋₄₅ Or ₂₋₃)	Antiperthitic andesine (An ₄₈₋₂₃ Or ₆₋₂₅)
Ferromagnesian minerals	Augite, bronzite, olivine	Bronzite-hypersthene
Iron-titanium oxide minerals	Magnetite + ilmenite	Hemo-ilmenite
Predominant rock type	Gabbroic anorthosite	Anorthosite
Shape	Irregular	Domical
Contact relations	Obscure	Intrusive
Associated rocks	Varied	Pyroxene syenite gneiss and pyroxene granite gneiss
Age	?2 Ga	Mostly in areas of 1 Ga old heating
Relation to other type	Contains dykes of andesine anorthosite	Contains xenoliths of labradorite anorthosite

Source: Anderson and Morin, in Isachsen (1969, p. 58).

normal rock being about An₄₆. They are most abundant in the Grenville Province of Canada (Fig. 3) but are widespread throughout the world (Anderson, in Isachsen, 1969).

In those with igneous textures a very coarse grain size is common; some plagioclase crystals may be measured in meters. The accompanying ferromagnesian minerals may show a variety of textures, some poikilitically enclosing plagioclase, others being themselves enclosed. "Protoclastic" texture, in which large grains of feldspar are broken down into smaller equidimensional grains, is another typical texture, although its significance is still debatable. In metamorphosed

masses the ferromagnesian minerals frequently occur as lenses in a zonal arrangement, with garnet as the outermost zone.

The origin of massif-type anorthosite presents great problems (Isachsen, 1969; Emslie, 1978; Weibe, 1980; Morse, 1982; Windley, 1984). There now seems no doubt that the major masses are igneous plutons, intruded under a variety of conditions. Whether they are differentiates from a basaltic parent or from dioritic magma is still unresolved and the formation of anorthositic melts by partial melting of lower crust has been suggested by various authors. Windley (1969) suggests that the middle Precambrian plutons are derived from an early Precambrian lower crust rich in Fiskenaasset-type anorthosites; Michot and Michot (in Isachsen, 1969) recognize several types of anorthosite in southern Norway at least one of which is formed by anatectic melting of crustal rocks; Moorlock et al. (1972) describe partial melting of basic and intermediate granulites to give anorthosite migmatites and cross-cutting narrow dykes of hypersthene anorthosite and biotite anorthosite that probably crystallized at very high temperatures and pressures. Geophysical evidence suggests that there are no "hidden" basic masses beneath anorthosite plutons (Philpotts, in Isachsen, 1969) so that differentiation from normal basic magma seems to be unacceptable, although some masses do show layering and ferrogabbro, norite, and troctolite masses are frequent associates even when large acidic and alkaline

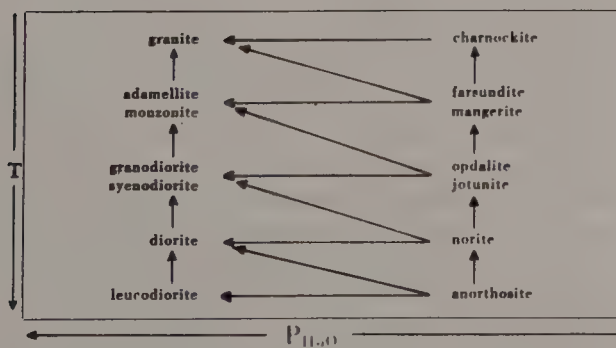


FIGURE 4. Relationships between the anorthosite-charnockite series and the leucodiorite-granite series; arrows indicate possible transitions between members of each series as water pressure changes and temperature falls. (From de Waard, in Isachsen, 1969, p. 77.)

bodies are also present. It seems most probable that in early and middle Precambrian times crustal conditions favored the formation of hydrous anorthositic liquids in much the same way that granitic magmas formed beneath orogenic belts at later periods although, of course, granitic magmas were also being formed contemporaneously with the anorthositic magmas.

Model constraints interpreted from mid-ocean ridge regimes may play a significant role in elucidating the development of Proterozoic magma systems and Flower (1984) presents a mid-ocean ridge model for anorthosite genesis. It invokes iterative plagioclase accumulation, magma mixing and phenocryst resorption to produce plagioclase-supersaturated parent liquids. Periodic replenishment of reservoirs in an environment of incipient or very slow dilation like that operative at mid-oceanic ridge systems is considered likely to produce the conditions for massive anorthosites to form.

Metamorphic-stratigraphic type

The metamorphic-stratigraphic type was first recognized by Harpum (see Windley, 1969), who termed it the Sakeny-type, after the Sakenites of Madagascar (Lacroix, see Windley, 1969). These anorthosites are found as "conformable stratigraphic" units in gneisses, usually of very high grade.

Because of the very strong deformation and multiple periods of high-grade metamorphism, to which most of these anorthosites have been subjected, it is not surprising that their origin is normally obscure. Because of its extreme chemistry, anorthosite, with rock types such as quartzite, marble, graphite schist, chromitite, and peridotite, resists the homogenization process that converts most rocks to granodioritic gneiss during high-grade regional metamorphism. Of these extreme rock types, anorthosite, chromitite, and peridotite are all original constituents of alpine-type layered complexes and also of the Fiskenaesset-type complexes of the early Precambrian. When these become involved in later orogenies, anorthosite lenses and bands tend to be preserved with accompanying chromite and ultramafic bands as units of the gneiss. The known examples (e.g., Sakeny, Malagasy; Sittampundi, S India; Fiskenaesset, Greenland) all have plagioclase of extreme calcic composition but only in the Fiskenaesset Complex is the original igneous origin unequivocal. High-grade metamorphism seems to result in plagioclase of almost pure anorthite composition and the rocks commonly have almost no Na. Other minerals occur (zoisite, corundum, spinel) and where present in

abundance, these normally indicate that desilication has taken place, e.g., the highly decorative Tanganyika Artstone from Merkerstein, N Tanzania, is a zoisite-ruby-corundum rock probably derived by desilication of anorthosite bands in a migmatite—this may be regarded as a meta-anorthosite although its proportion of anorthosite is very small.

Proposals that sedimentary layers were transformed into anorthosite are not widely accepted. Heitanen (in Isachsen, 1969) suggests small-scale metasomatic alteration of sedimentary shale and limestone layers as the origin of anorthosites in Idaho and here the evidence of alteration along the layers to rocks of more normal type is convincing and the complete lack of any other metamorphosed igneous rock types supports this interpretation. It seems, however, to be a very uncommon type of reaction.

Other terms

Anorthitite—an igneous rock almost completely composed of anorthite; also referred to as an anorthite rock.

Bytownitite—an igneous rock composed almost entirely of bytownite.

Labradophyre—an anorthosite composed of labradorite phenocrysts in a groundmass of the same mineral.

Labradoritite—a rock composed almost entirely of labradorite.

Modumite—an anorthosite with plagioclase of bytownitic composition and some pyroxene and barkevikite.

Plagioclasite—anorthosite.

A. E. WRIGHT

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Cross-references: *Alpine mafic magma stem; Bushveld Complex; Charnockite suite; Double-diffusive convection; Gabbro and norite; Intrusion-layered; Lunar petrology; Plutonic rocks—classification and nomenclature; Skaergaard intrusion; Ultrabasic igneous rocks.*

APLITE

Aplite is a medium- to fine-grained rock (average grain size usually 1.0 mm or less) with a characteristic equigranular, sugary (saccharoidal) texture. The origin of the term is uncertain, but the name may have been used as early as 1800. In 1823 it was used by von Leonhard for medium-grained granitic rocks that lack mica and are composed essentially of quartz and feldspar. The name, derived from the Greek *απλόος*, simple, alludes to this. Most early petrographers followed von Leonhard's usage though H. Rosenbuch widened the term to include most leucocratic dyke rocks. There are aplites corresponding in composition to many of the principal rock groups, for example, granite aplite, diorite aplite and so on. When the name is used without a prefix it is generally taken to refer to granite aplite.

The principal constituents of granite aplite are quartz and alkali feldspar (usually orthoclase, microcline, perthite, or sodic plagioclase) with

minor proportions of muscovite, biotite, or hornblende. Quartz and K-feldspar are frequently micrographically intergrown, and aplites may also contain topaz, fluorite, garnet, tourmaline, and axinite as accessory minerals.

Aplites occur typically as dykes, veins, and stringers, usually less than a meter in thickness, and with sharp or gradational contacts. They are emplaced either within the parent igneous body or in the country rocks at no great distance from it. They may be banded parallel to their margins and are commonly associated with pegmatite so that the two may be interlayered or occur as patches one within the other.

Most granite aplites appear to have crystallized from siliceous residual solutions derived from granitic magma. Aplite and pegmatite are considered to be the volatile-poor and volatile-rich parts, respectively, of these residual solutions and this interpretation accounts in some way for their close association. Some aplites, particularly those occurring in metamorphic terranes, are of replacement origin and recrystallization of country rock to form aplite may have been facilitated, in some instances, by shearing.

The following are varieties of aplitic rocks.

Akenobeite—a granodioritic aplite (Akenobe district, Japan).

Beresite—a quartz-rich variety of aplite, altered to material resembling greisen and often characterized by the presence of pyrite (Beresovsk, Urals, U.S.S.R.).

Bostonite—a light-colored, aplitic, alkali syenite composed largely of albite and microcline; the laths or microlites of feldspar, in part showing a crudely radial arrangement and in part being more or less parallel, in a trachytoid groundmass, give the rock a bostonitic texture (Boston, Massachusetts, U.S.A.).

Brandbergite—a hypabyssal rock with an aplitic texture and composed of orthoclase, quartz and aggregates of biotite in a micrographic groundmass.

Gladkaite—a gray hypabyssal rock composed of a granular aggregate of sodic plagioclase, quartz, hornblende and some biotite, sometimes grouped as a quartz lamprophyre; with disappearance of quartz, the rock passes into a plagiopelite (Gladkaia Spoka, U.S.S.R.).

Krageröite—a rutile-bearing albite aplite with minor proportions of quartz, K-feldspar, and ilmenite (Kragerö, Norway).

Lestivarite—a syenite-aplite or microsyenite composed chiefly of microperthite, acmite, arfvedsonite, and accessory sphene (Lestivare, Finland).

Maenaite—a hypabyssal calcic bostonite, intermediate between a monzonitic felsite and a

dioritic lamprophyre, and characterized by an abundance of hornblende (Maena, Norway).

Plagioplite—a dioritic aplite composed chiefly of plagioclase (oligoclase to andesine) with subordinate green hornblende and micas.

A. C. BISHOP

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Cross-references: *Banding in igneous rocks*; *Granite*; *Granites and mineralization*; *Pegmatite*; *Volatiles*.

APPINITE

The *appinite suite*, which occurs in pipe-like intrusions in the British Caledonides, includes ultrabasic-ultramafic (peridotite, pyroxenite, hornblendite), basic to intermediate (kentallenite and many varieties of diorite), and acidic (granodiorite) types. Abundant amphibole is a common characteristic. A cluster of intrusions in the parish of Appin in W Scotland (Fig. 1) formed the basis for the term appinite (Bailey and Maufe, 1916) and for the demonstration of the subvolcanic characters of the masses and their association with explosion-breccia (Fig. 2). Another cluster 50 km to the SE includes the Garabal Hill intrusion, which Nockolds (1941) used as a basis for proposing that the spatially-related and much more voluminous granitic plutons were genetically related to the appinite suite. Small clusters and isolated pipes occur in other parts of Scotland and a major cluster occurs in County Donegal, Eire (Fig. 7). Here, as elsewhere, there is a spatial association of appinites with plutons of granodiorite (Pitcher and Berger, 1972). Rocks of the appinite suite are not common in the Scandinavian or the North American extensions of the Caledonides but igneous complexes of similar affinity are known from eastern Australia and they are widespread in Tanzania where there are associated lamprophyres and breccia bodies.

Appinite was defined by Baily and Maufe (1916) as the plutonic equivalent of vogesite and spessartite, i.e., of the hornblende-rich lamprophyres, rocks that are abundant throughout the British Caledonides. Together with these and other lamprophyric rocks, the appinites include representatives of dominantly or wholly mantle-derived (primary) magma compositions.



FIGURE 1. Distribution of Caledonian igneous masses, including rocks of the appinite suite, in W Scotland.

Rocks of appinite suite

Masses of cortlandtite, biotite pyroxenite, biotite hornblendite, pyroxene diorite, pyroxene-mica diorite, kentallenite, monzonite, quartz diorite, syenite, and granodiorite occur as composite bodies together with hornblende diorite and appinite in the type region. Elsewhere scyelite (mica-hornblende peridotite) occurs. Gradation from hornblendite to appinite and from appinite to lamprophyre forming marginal facies is seen (Fig. 3a). Other amphibole-rich varieties consist of large pargasite phenocrysts in a lamprophyric matrix.

With the existence of such variations, some uses of the term appinite have been related to the overall rock association. However, when used in relation to a specific rock type, appinite has been used to designate a rock in which large idiomorphic pargasitic amphibole dominates in a mafic rock of gabbroic to dioritic affinities (Fig. 3b). Some very coarse-grained appinites are of pegmatitic aspect. Commonly, the large amphiboles contain corroded diopside and other minerals, while they have partly corroded outlines and rhythmic amphibole overgrowths of tremolite and actinolite (Fig. 3c). Rarely, the

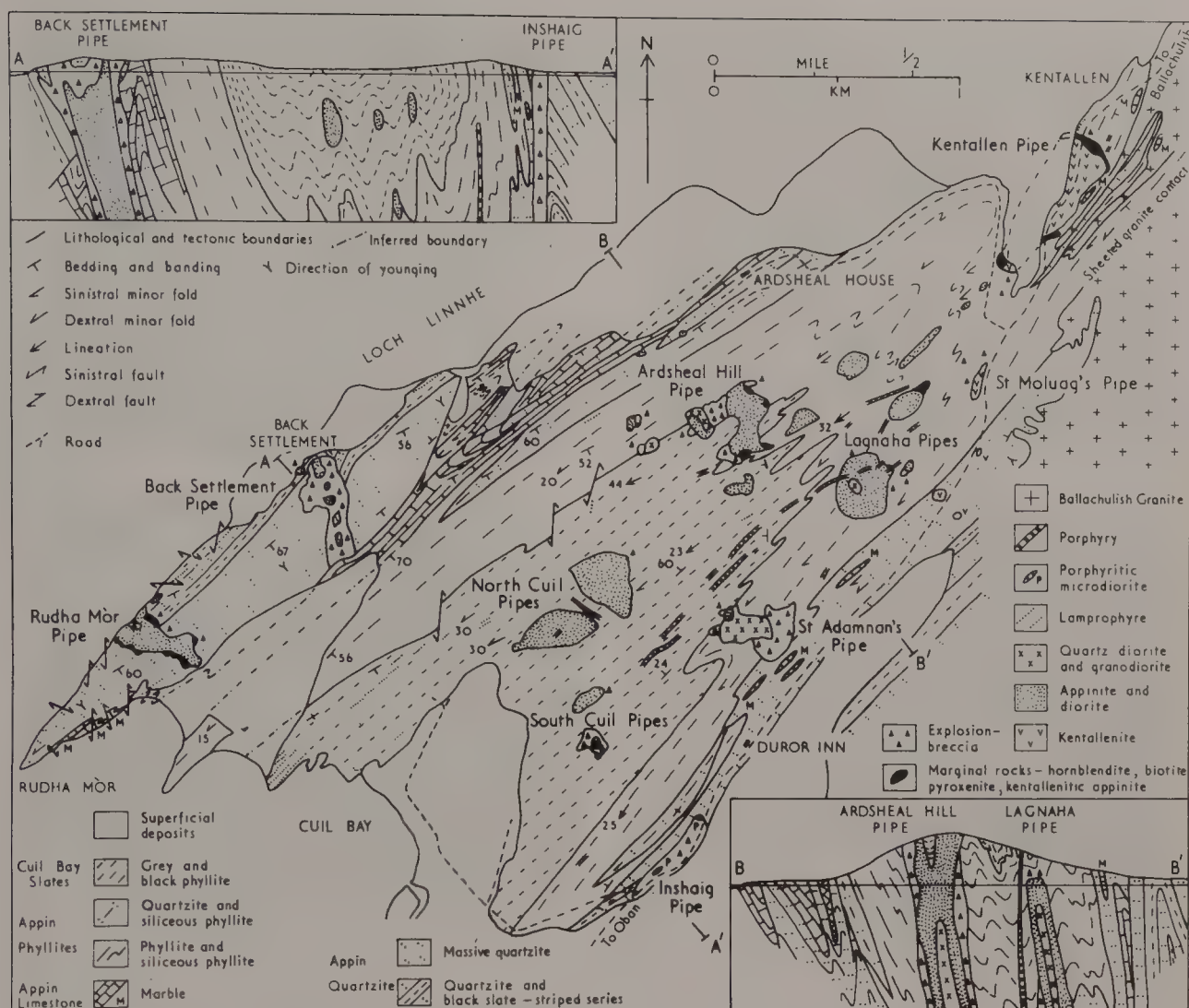


FIGURE 2. Explosion-breccia-appinite pipes in the area near Kentallen, W Scotland. (After Bowes and Wright, 1967.)

idiomorphic pargasite crystals show spectacular zoning (Fig. 3d). In associated rocks, clinopyroxene can show comparable zoning and biotite comparable overgrowths.

Despite the wide variations in mineralogy, geochemical studies have shown the suite to represent the products of volatile-rich basaltic magma of high-K calc-alkaline type (Table 1). Most crystallization and magmatic differentiation took place in pipe-like intrusions that represent potential feeder channels for lava flows (Figs. 1, 2) that were impeded by structural traps (Fig. 4; Bowes and Wright, 1967). Other theories proposed to account for the petrogenesis of the appinite suite include (1) the hybridization of solid ultramafic rocks or ultramafic magma, (2) limestone-granite reaction and crystallization from a magma of pyroxene-mica diorite composition (see Bowes and McArthur, 1976, and references therein).

Conditions of crystallization

On the basis of mineral chemistry, Hamidullah and Bowes (1988) showed that olivine was the first mineral on the liquidus and crystallized at a temperature of ca. 1,800°C under a pressure of ca. 25 kb (depth 70–80 km). Pressure estimates for much of the suite, and the occurrence of explosion-breccia, indicate subsequent emplacement at high crustal levels. There the bulk of crystallization took place under decreasing temperature conditions but highly variable P_{GAS} in subvolcanic pipes. CMAS plots show that there was successive fractionation of clinopyroxene, amphibole, mica, and feldspar. Clinopyroxene crystallized in the temperature range of 1,125–1,000°C, while $P_{\text{H}_2\text{O}}$ was increasing from <1 kb to 1.5 kb (Fig. 5). With increasing $P_{\text{H}_2\text{O}}$, resulting from gas escape being impeded by structural traps,

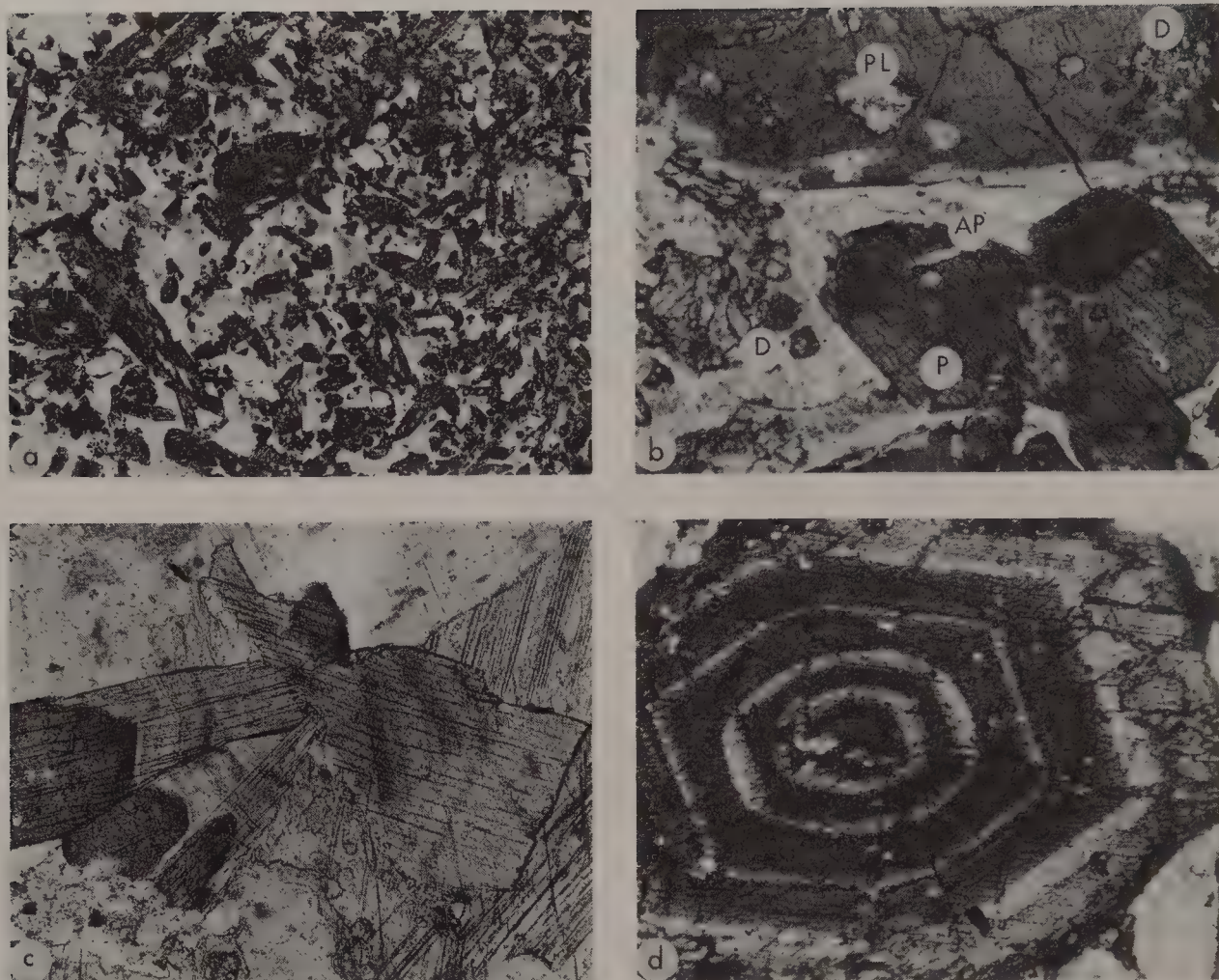


FIGURE 3. Photomicrographs of characters of appinites from the Appin district, W Scotland. (a) Gradation from appinite (left) to lamprophyre; both rocks composed essentially of hornblende and feldspar; PPL, $\times 4$. (b) Appinite consisting of pargasite (P) with inclusions of diopside (D), albite (PL), and apatite (AP) in a groundmass of diopside and albite; PPL, $\times 20$. (c) Rhythmic amphibole overgrowths of actinolite and tremolite on corroded hornblende crystals surrounded by plagioclase and calcite; PPL, $\times 11$. (d) Zoned pargasite, XN, $\times 23$.

the large idiomorphic pargasite crystals (Fig. 3b) formed at $960\text{--}770^\circ\text{C}$ and $P_{\text{H}_2\text{O}}$ of 2–4 kb. Accumulation of mafic phases occurred while the pyroxene and amphibole were crystallizing, the latter ending with the development of actinolite and tremolite overgrowths in a temperature range down to 750°C but with $P_{\text{H}_2\text{O}}$ above 5 kb, but varying up and down.

Explosive activity associated with the final breaching of the traps led to the disruption of early-formed cumulates and the upward (explosive) movement of crystal mushes and individual crystals with variable proportions of residual liquid. This led to many phenocrysts not being in a matrix representing the particular liquid in which they were in equilibrium. The result was both the production of an even greater variety of mineral assemblages and textural variants than produced by crystallization differentiation and rocks that contain mixtures of minerals formed at different

crystallization stages, both before and after the explosive activity.

The late leucocratic diorites and granodiorites largely represent the residual melt: these crystallized at $<770^\circ\text{C}$ and at ca. 1 kb $P_{\text{H}_2\text{O}}$. They probably contain a component of a secondary liquid, produced by solution (under elevated P_{GAS} conditions) of silica from quartzite that formed the structural traps to the rising magma. Subsequent gas-streaming led to hydration and carbonation of some of the products of magmatic crystallization.

Magma type and source

Geochemical characteristics of the rocks from the type area for appinites, including the continuous differentiation shown on an AFM diagram (Fig. 6), Ni vs. SiO_2 variation, and increasing Sr in the residual liquid, indicate an affiliation of this appinite assemblage to the

TABLE 1. Mean major and trace element analyses and modal compositions of rocks of the appinite suite and lamprophyres, Appin district, W Scotland

	1	2	3	4
Major elements				
SiO ₂	45.10	48.31	48.89	49.00
TiO ₂	1.23	0.97	0.63	0.76
Al ₂ O ₃	10.59	12.71	12.53	13.82
Fe ₂ O ₃	3.09	2.60	2.11	2.07
FeO	6.78	6.63	6.41	6.32
MnO	0.14	0.13	0.12	0.13
MgO	17.35	11.60	14.84	8.62
CaO	8.65	8.40	7.19	7.73
Na ₂ O	1.67	2.44	2.56	2.79
K ₂ O	2.51	2.13	2.20	2.14
P ₂ O ₅	0.06	0.25	0.28	0.28
H ₂ O + CO ₂	2.17	2.65	2.02	3.11
	0.49	1.31	0.43	3.29
TOTAL	99.83	100.13	100.21	100.06
Trace elements (p.p.m.)				
Cr	758	453	759	316
Co	70	43	56	46
Ni	357	160	430	126
Ba	1025	899	770	825
Ce	32	37	46	48
Cl	439	263	422	241
Cu	43	107	83	121
Ga	7	11	8	12
La	10	43	41	34
Li	11	10	7	14
Nb	7	8	5	5
Pb	8	15	20	24
Rb	58	44	61	57
S	638	868	205	1913
Sc	55	43	27	34
Sr	411	833	965	942
Th	3	5	4	6
V	301	223	189	222
Y	14	16	14	15
Zn	62	93	73	89
Zr	55	87	86	102
Modes				
Ol	2	0	15	5
Cpx	60	20	40	17
Amph	10	45	0	18
Phl-bio	20	5	10	5
Plag	5	15	20	37
Orth	0	3	0	1
Anor	0	0	10	0
Qtz	0	4	0	5
Cal	0	4	0	3
Opaq	2	1		6
Others	1	3	1	3

Source: from Wright and Bowes (1979) and Hamidullah and Bowes (1988). Ol, olivine; Cpx, clinopyroxene; Amph, amphibole; Phl-bio, phlogopite-biotite; Plag, plagioclase; Orth, orthoclase; Anor, anorthoclase; Qtz, quartz; Cal, calcite; Opaq, opaque ores.

1, biotite pyroxenite (cf. Fig. 5a).
2, appinite (cf. Fig. 5b).
3, kentalenite.
4, lamprophyres and chilled margins of appinitic intrusions.

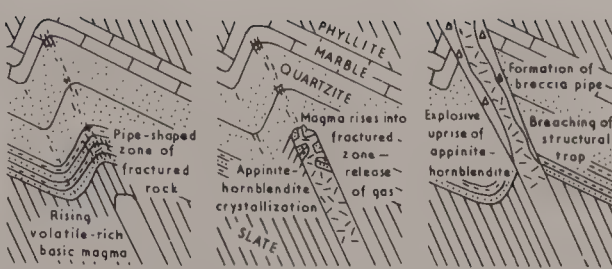


FIGURE 4. Representation of stages in the formation of explosion-breccia-appinite pipes. (After Bowes and Wright, 1967.)

calc-alkaline suite. However, with its higher K₂O proportion than for the normal calc-alkaline series, a high-K calc-alkaline series is indicated. For the appinite assemblage in the Loch Lomond district (Fig. 1), derivation from a normal calc-alkaline magma is indicated by its geochemical characteristics, while for the appinite suite of Donegal, Eire, the magma type was transitional between calc-alkaline and tholeiitic (Fig. 7). Such variations correspond with variations in modern-day calc-alkaline

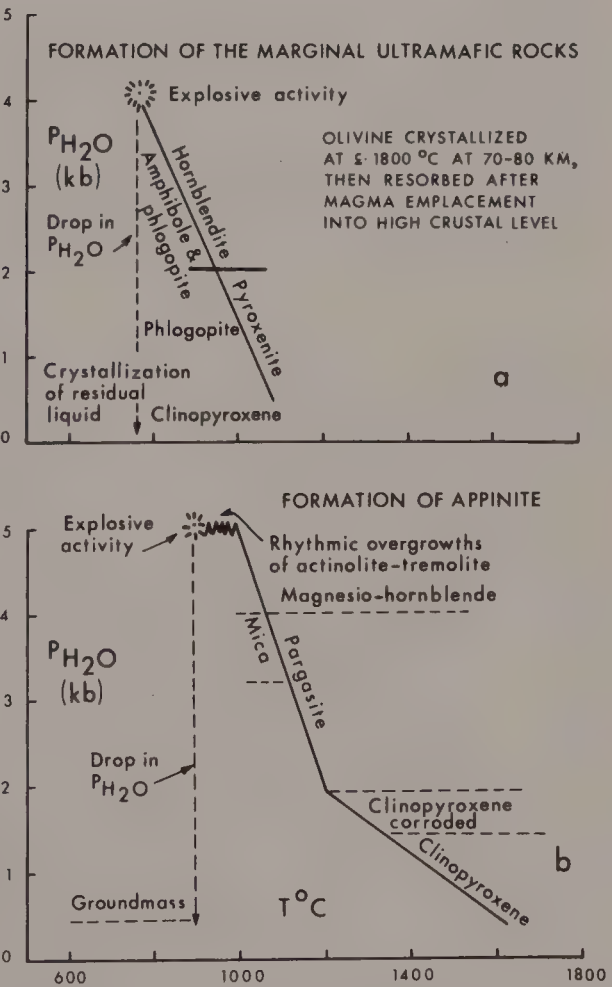


FIGURE 5. Schematic representation of P-T conditions for igneous and explosive activity for (a) marginal ultramafic rocks and (b) appinite. (After Hamidullah and Bowes, 1988.)

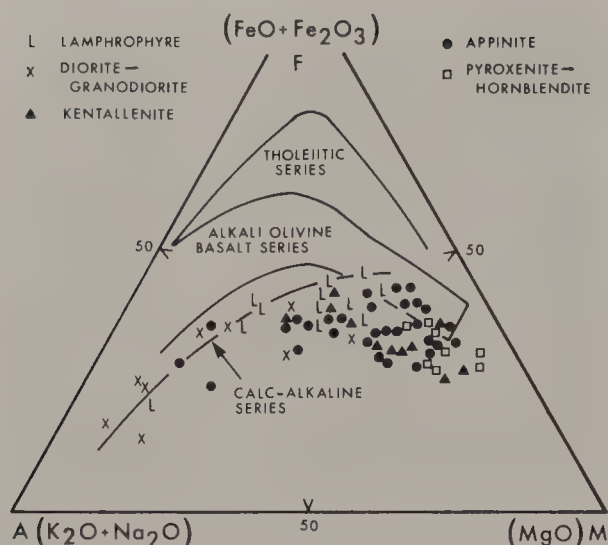


FIGURE 6. AFM (($K_2O + Na_2O$)–($FeO + Fe_2O_3$))–MgO plot of rock compositions of the appinite suite of the Appin district, W Scotland, with trends of various basalt series. (After Hamidullah and Bowes, 1988.)

volcanic suites associated with subduction. Accordingly, a major factor in the formation of the various representatives of the appinite suite in the British Caledonides is interpreted as being a NW–WNW-dipping subduction zone associated with the closure of Iapetus ocean (Fig. 7). Using the determined temperature of formation of the olivine (ca. 1,800°C) and an average Phanerozoic geothermal gradient, derivation at minimum depth of 200 km is deduced

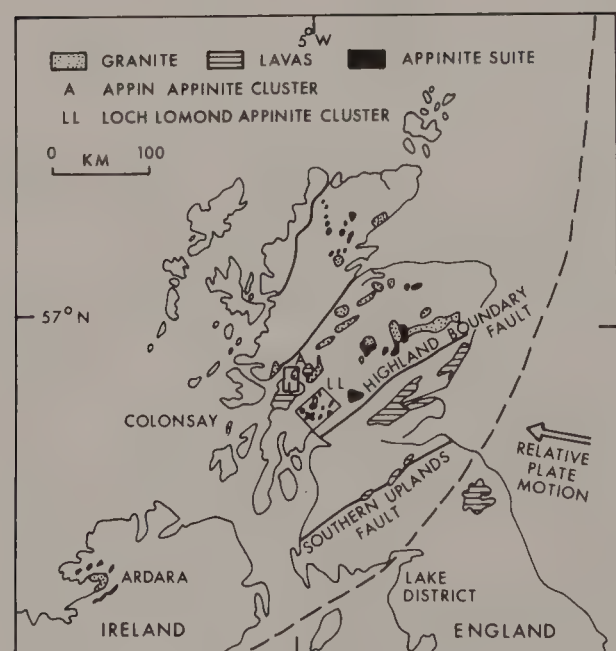


FIGURE 7. Distribution of the appinite suite, lavas, and Newer granites of the late Caledonian magmatism in the British Caledonides and position of the late Silurian NW–WNW-dipping subduction zone.

for the basaltic magma from which the appinite suite in its type area was derived.

The uprise of mantle-derived basic magma took place at a time when the lower crust in the Caledonian orogen was the site of anatexis activity that culminated in the emplacement of the voluminous late Caledonian granitic masses. The spatial association of clusters of appinitic pipe-like masses and granitic plutons could reflect uprise of the basic magma as mantle diapirs. Their heat could have been a major factor in the development of large amounts of crustal melt. In addition, differential “scouring” of elements from a lower crust in an anatexis state by uprising gas-rich basic magma could explain apparently anomalous geochemical features of the appinite suite (e.g., high K, Rb, and Ba with high Ni and Cr—Wright and Bowes, 1979).

D. R. BOWES

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Cross-references: Breccia pipe; Calc-alkaline rocks; Diatreme; Durbachite; Lamprophyre; Magmatic differentiation; Plutonic rocks—classification and nomenclature; Volatiles.

ARCHAEOAN VOLCANISM

The products of *Archaean volcanism* are a major component of Archaean greenstone belts that are present in most Precambrian shields of the world and that show worldwide similarities in first-order features despite their many

* Contains extensive list of references.



FIGURE 1. Distribution of volcanic rocks in greenstone belts of the Canadian Shield.

significant and fundamental points of difference when considered in detail. Being repositories of major mineral deposits, particularly of Au, they have received wide attention.

Archaean greenstone belts are typically linear to irregular-shaped, synformal supracrustal successions, 5–250 km wide and up to several hundred kilometers long (Fig. 1). Exposed thicknesses are 10–20 km, although true stratigraphic thicknesses are often uncertain. Most belts represent faulted synforms with fold axes and major faults parallel to synformal axes. The typical keel-shaped outline is partly the product of diapiric intrusive plutons. Metamorphic grade is typically greenschist but may be up to granulite facies; it may also be higher near intrusive contacts. Primary textures and structures are often remarkably well-preserved in the metasupracrustal rocks. The greenstone belts typically form integral but minor (20%) components of widespread granitoid-greenstone terranes in which the major components are gneissic complexes, diapiric intrusions, batholiths, and late discordant plutons (Fig. 2). Although worldwide the common ages of greenstone belts are 2.6–2.7 Ga, ages to 3.5–3.8 Ga are well represented.

In the Canadian Shield studies were conducted by A. C. Lawson and others in the latter part of the 19th century. There the general physical characteristics of Archaean volcanic rocks were recorded and gross similarities to modern volcanic rocks established in the first half of the 20th century with J. W. Ambrose, H. C. Cooke, H. C. Gunning and M. E. Wilson

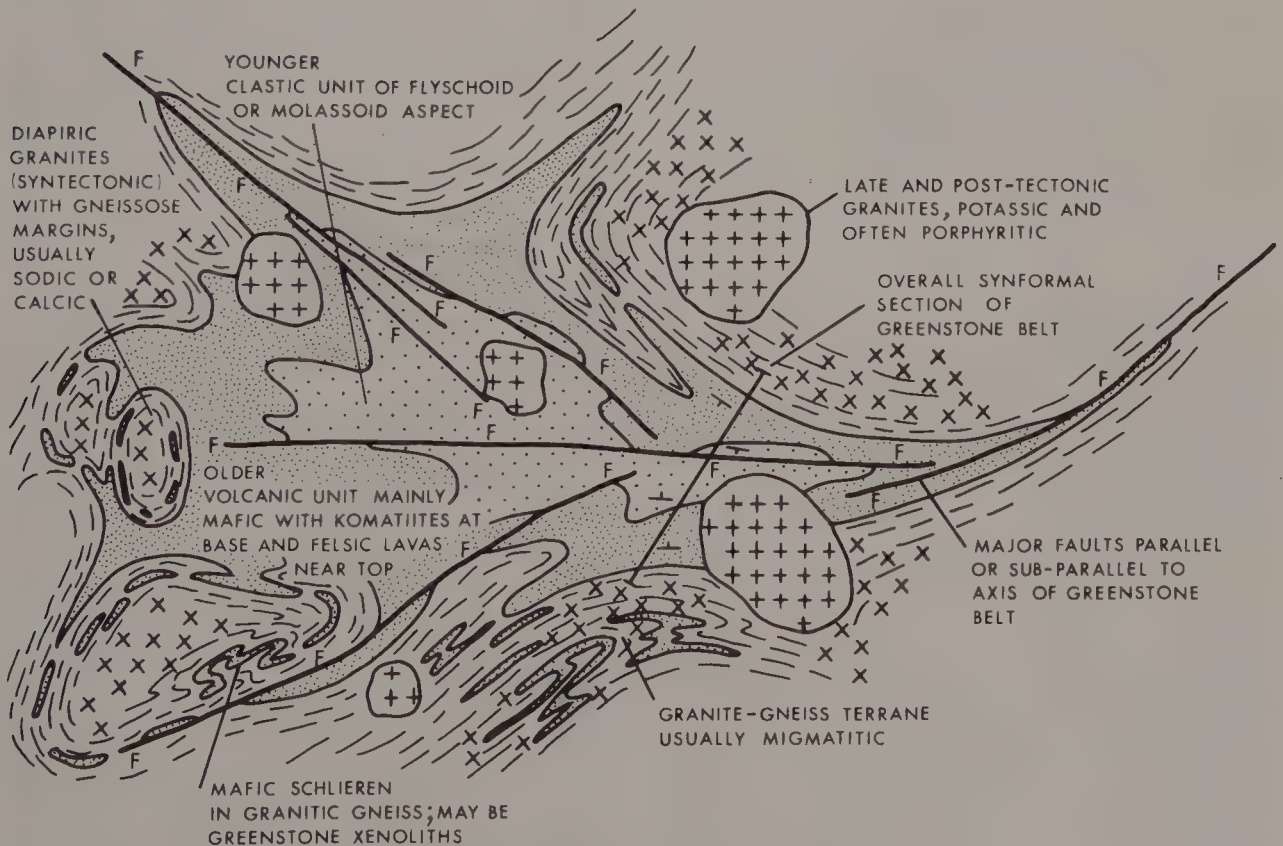


FIGURE 2. Schematic representation of a greenstone belt.

playing prominent roles. Accelerated field and laboratory studies especially in Canada, southern Africa, India, and Australia during the past 30 years have shed much light on Archaean volcanic processes. One important result has been to establish a gross similarity to processes operating in the development of Cenozoic orogenic belts, particularly those at continental margins and notably in island arcs. However, significant points of difference between Archaean greenstone belts and Phanerozoic volcanic arcs are (1) much smaller length-to-width ratios, (2) near-absence of glaucophane schist (blueschist) facies metamorphism, and (3) abundance of ultramafic lavas in the Archaean units. Another major problem concerns the uncertain nature of the basement upon which Archaean greenstones were formed. Despite examples of undoubted unconformities such as at the base of the Dharwar greenstone belts in Karnataka subprovince, India, and the Belingwe greenstone belts in the Shabani area, Zimbabwe, greenstone belt contact relations are typically equivocal, being either intrusive-metamorphic or faulted. As a result, the ensimatic or ensialic basement relationships can seldom be established by direct observation.

Because of the unusual size and variety of greenstone belts and the long history of study, those of the Canadian Shield are mainly used here to illustrate the nature of Archaean volcanism worldwide. Information about Archaean volcanism in greenstone belts elsewhere is given by Condie (1981) and Windley (1984) and references therein. Anhaeusser and Wilson (1981) and Hallberg and Glikson (1981) give major summaries for southern Africa and Australia, respectively.

Products

Archaean volcanic products, in common with those of modern volcanism, take the form of both solidified lava flows and fragmental forms (Goodwin, 1977; Dimroth et al., 1982). Although some flows are massive throughout, most include massive, pillowed, and breccia phases. Commonly, these three phases are arranged in ascending order with lower massive, upper pillowed, and uppermost breccia phases. Pillow structures are widespread and indicate a prevailing subaqueous environment of accumulation. Fragmental forms include ejected blocks, cinders, ash, and dust as well as hyaloclastic and pyroclastic deposits. Globular or ellipsoidal bombs with glassy rims, now devitrified, and scoriaceous interiors are quite common. Many fragmental layers display graded bedding as a result of settling in air and/or water. The products of violent explos-

ion (*nuées ardentes*) have been positively identified in only a few Archaean assemblages. This may reflect a comparative paucity of subaerial eruption and preservation in Archaean time and the corresponding dominance of subaqueous eruptions. In keeping with this situation, however, submarine ash flows derived by subaqueous pyroclastic eruption are common.

Basalt and andesite lava flows commonly feature amygdules, variolites, scoria, metapalagonite, pillow structures, columnar joints, and ropy and blocky flow surfaces. In more felsic lavas spherulites, amygdales, perlitic cracks, and lithophysae are common. Although felsic volcanic rocks are commonly in fragmental form owing to explosive expulsion, lava flows do predominate locally as at Noranda, Quebec. In such cases, the rhyolite and dacite lava flows display progressive fragmentation ranging from massive lava near the vent through a partly brecciated phase with large ball-like structures in the intermediate distances to completely fragmented forms in the distal part of the flow (Gélinas and Ludden, 1984).

Although the products of Archaean volcanism have been varyingly altered and destroyed through processes of deformation and metamorphism, it is clear that they were originally distributed in the form of extensive lava fields or plateaus as well as large individual volcanoes, including both shield volcanoes formed of predominant lava flows, and, more commonly, composite volcanoes composed of alternating lava flows and pyroclastic layers. Delineation of individual and mutually coalescing neighboring volcanoes is possible in places (Dimroth et al., 1982; Goodwin, 1982).

Lithology and stratigraphy

Lava flows and fragmental products commonly of the komatiite-basalt-andesite-dacite-rhyolite association are typically arranged in mafic to felsic cycles or successions (Goodwin, 1982). The volcanic rocks are intercalated, especially in upper stratigraphic parts, with volcanogenic sediments, mainly graywacke, mudstone, tuff, conglomerate, and cherty iron-formation. Subvolcanic granitoid plutons are common. Mafic to ultramafic intrusions in the form of discrete sills, dykes, and small irregular plutons are widespread although in small proportions (Fig. 2).

The flows and pyroclastic rocks are mainly tholeiitic to calc-alkaline in character but some alkalic rocks are present locally. Basalt lava flows with local ultramafic (including komatiitic) components predominate in lower stratigraphic parts. Andesite flows and pyroclastic rocks are

TABLE 1. Relative proportions of different types of volcanic rocks in Canadian Archaean volcanic belts

	Komatiite	Basalt	Andesite	Dacite	Rhyolite
Birch-Uchi	0.8	54.7	27.2	11.6	5.7
Lake of the Woods-Wabigoon	0.4	52.1	33.8	10.2	3.5
Timmins-Kirkland Lake-Noranda	0.9	55.1	33.3	6.7	4.0
Superior Province—average	0.7	54.0	31.4	9.5	4.4

intercalated with basalt and, in many assemblages, increase proportionately upwards to a majority position in intermediate to upper levels of the assemblage. Felsic rocks, mainly of pyroclastic type but locally massive and fragmental lava flows, are generally present in the upper stratigraphic parts. Many Archaean assemblages display a single generalized mafic to felsic succession; some have, in addition, a mafic capping; still others display two or more superimposed mafic to felsic cycles. The common stratigraphic thickness is in the order of 12,000 m.

Abundance of volcanic rocks

Table 1 shows the relative proportions of different types of volcanic rocks in three Archaean volcanic belts of the Canadian Shield (Goodwin, 1977). There are markedly higher proportions of ultramafic-mafic rocks in the Barberton (South Africa) and Zimbabwe greenstone belts compared to those in Canada (Anhaeusser and Wilson, 1981). Proportions of ultramafic-mafic rocks:basalt + andesite + felsic volcanic rocks are 1:3.1 and 1:9.2 for the Barberton and Zimbabwe terranes, respectively, and 1:19.8 and 1:25.3 for the Abitibi and Wabigoon belts, Canada, respectively. In the case of the Barberton terrane, proportions of ultramafic rocks:basalt:felsic volcanic rocks are 6.6:10.5:1; for the tholeiitic association in modern island arcs, basalt:andesite:“dacite” are 50:35:15 (Jakeš and White, 1971). Hence, although Archaean assemblages contain more basalt and komatiite and less andesite than modern arcs of tholeiitic association, their general similarities are remarkable and suggest

similar histories of origin and tectonic development.

Compositions

The weighted average compositions of three representative volcanic belts in Superior Province, Canada, show remarkable similarities (Table 2). They also correspond closely to the average composition of a “developed island arc” (Jakeš and White, 1971) and even more closely to that of a more primitive or tholeiitic (young) island arc (the oceanward part of a fully developed island arc, e.g., South Sandwich Is., Central Kuriles). The main distinguishing characteristics are significantly higher FeO, Ni, V, Cr, and CO₂, and somewhat lower Na₂O and CaO in Archaean rocks.

Weighted average compositions for six volcanic classes present in a representative Archaean volcanic belt are given in Table 3. These are based on massive lava flows mainly, but also on fragmental products, with the determination of class based on the classification of Irvine and Baragar (1971). Based on this and other data, it is evident that the volcanic rocks of different Archaean belts are derived from the same or similar magmas and underwent similar processes of differentiation.

Comparison with modern island arcs

Archaean volcanic belts present close analogies to modern island arcs in terms of (1) prevailing mafic to felsic stratigraphic successions, (2) similar chemical compositions and abundances of volcanic classes, (3) widespread calc-alkaline volcanism including much highly

TABLE 2. Weighted average chemical composition of volcanic belts in the Superior Province, Canada

Major elements (wt.%)														
Volcanic belt	N	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	H ₂ O	CO ₂
Birch-Uchi	116	54.3	0.83	14.5	2.62	7.06	0.16	4.93	6.84	2.70	0.67	0.26	—	4.56
Lake of the Woods	420	54.3	0.87	14.9	2.37	7.30	0.17	4.81	7.25	2.78	0.75	—	1.98	2.06
Timmins-Noranda	1080	53.7	1.07	15.1	2.21	6.93	0.18	4.96	6.92	3.27	0.74	0.16	2.93	1.17
Average		54.1	0.92	14.8	2.40	7.10	0.17	4.90	7.00	2.92	0.72	0.21	2.46	2.60

TABLE 3. Weighted average compositions of six volcanic classes present in the Lake of the Woods—Wabigoon region, western Superior Province, Canada

Major elements (wt.%)														
Volcanic class	N	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	H ₂ O	CO ₂
Komatiite	5	38.7	0.60	4.32	7.04	7.79	0.19	27.36	3.77	0.28	0.04	—	3.65	7.17
Basalt	155	49.2	1.01	14.3	2.79	9.12	0.21	6.36	8.97	2.27	0.35	0.27	2.27	2.38
Andesite	146	55.9	0.89	15.3	2.31	6.39	0.15	4.02	5.97	3.40	0.92	0.31	1.75	1.43
Dacite	79	65.6	0.49	15.2	1.37	3.30	0.09	2.05	3.59	4.10	1.41	0.26	1.01	1.30
Rhyodacite	9	65.9	0.57	15.3	1.37	3.60	0.17	1.72	2.82	2.63	2.38	0.12	1.40	1.62
Rhyolite	28	72.9	0.23	14.5	1.20	1.29	0.07	0.68	1.20	3.80	2.10	0.18	0.71	0.89

explosive activity associated with felsic masses, (4) predominantly subaqueous accumulation of the lava flows and sedimentary associates, and (5) parallel arc-like arrangement of successively younger linear volcanic belts (cf., Capdevila et al., 1982). However, the high proportion of komatiites in Archaean volcanic belts (Sun and Nesbit, 1978) points to a distinctive Archaean mantle composition and operating tectonic process. These relationships suggest a comparatively closely spaced interaction of “oceanic-type” and “island-arc type” volcanic material in the development of Archaean crust. Operating within a comparatively thin, supple, varied crustal regime the dominant tectonic process interpreted for Archaean time is aptly termed miniaturized “soft-plate” tectonism, as distinct from modern plate-tectonic activity that involves large, thick, rigid crustal plates. The resulting Archaean subsea-floor geothermal systems are closely linked to massive sulfide ore deposits (MacGeehan and MacLean, 1980).

A. M. GOODWIN

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Cross-references: *Basalt*; *Calc-alkaline rocks*; *Greenstone*; *Igneous rocks—tectonic setting*; *Island arcs—petrology*; *Nuée ardente*; *Spinifex texture*; *Ultramafic lavas*; *Volcanoes and volcanism*.

ARIÉGITE

Ariégite is a pyroxenite chiefly composed of clinopyroxene, orthopyroxene, and spinel, with pyrope and/or hornblende as possible accessories and lacking primary feldspar. The type rocks were described by Lacroix (1901, 1917), occurring as layers in lherzolite at Ariège, and elsewhere in the Pyrenees (Kornprobst and Conquéré, 1972). The particular rocks described by Lacroix have a mineral assemblage of diopside, bronzite, green spinel, and garnet. The term is still widely used in Europe, particularly by French geologists: in the IUGS nomenclature (Streckeisen, 1976) it would be termed diopside-bearing spinel pyroxenite or websterite (denoting a two-pyroxene rock). The absence of olivine distinguishes it from lherzo-

lite, and the composition of the pyroxenes distinguishes it from eclogite.

Kelyphitic reaction rims, especially around garnet, are the rule in ariégite, the rims consisting of complex intergrowths that may include spinel, anorthite, pyroxene, or amphibole. They result from the re-equilibration of the garnet with the pyroxenite groundmass and reflect the sensitivity of this assemblage to changing P_{TOTAL} , $P_{\text{H}_2\text{O}}$, and T conditions. In the same manner as with peridotites, experimental work has permitted the definition of P - T conditions for these rocks (O'Hara, 1967; Morse, 1980) which reflect a mantle origin.

In the type locality, ariégite forms segregations in the ultramafic lower portion of an ophiolite complex: it is found in a similar setting elsewhere. Like much of the lower ultramafic portion of such complexes, it typically displays a strong tectonic fabric (Nicolas et al., 1980). Rare ariégite nodules have been reported from kimberlite pipes and alkaline basalts, but the more usual association with harzburgite and lherzolite suggests that it represents a tectonized segregation formed during the generation of basaltic melts in the upper mantle beneath mid-ocean spreading ridges.

ADRIAN F. PARK

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- Cross-references: *Eclogite facies*; *Lherzolite*; *Pyroxenite*; *Reaction rim*; *Ultramafic rocks*; *Upper mantle—experimental petrology*.

ASSIMILATION

Assimilation is the process whereby solid or fluid foreign material is incorporated into

magma. The term implies no specific mechanism and the process depends on many factors, including temperature, cooling rate of the magma, gas pressure, and the composition of the material being assimilated. The result is a contaminated magma and, on cooling, rocks that are referred to as being contaminated. Wholesale assimilation requires the magma to be superheated, which is rarely the case for a fractionating magma. Hybridization can be synonymous with contamination but most commonly it refers either to mixing of two magmas or to one magma digesting earlier-formed magmatic material.

Reactions involved in the assimilation of sedimentary material by basic magma are given by Bowen (1928): magma crystallizing early-formed high-temperature members of the continuous and discontinuous reaction series incorporate the lower-temperature phases in the sedimentary material into the liquid fraction. High-temperature basic magma in equilibrium with olivine can incorporate refractory pelitic xenoliths containing phases such as cordierite and orthopyroxene (Gribble, 1968) while the presence of chromite layers in noritic rocks has been explained on the basis of incorporated siliceous material moving the path of crystallization (Irvine, 1975). Where, as in the case of granitic magma, material incorporated contains earlier-formed members of the reaction series than those crystallizing, complex reaction converts them to phases in equilibrium with the magma. Partly digested material commonly remains (as xenoliths).

The concept of alkaline igneous rocks being formed by desilication of basic magmas resulting from assimilation of limestones (Daly, 1910) is now not accepted, although, at high temperatures and on a small scale, considerable metasomatic changes can occur between limestones and igneous magmas with contamination occurring in the contact zone (e.g., at Scawt Hill, Northern Ireland: Tilley and Harwood, 1931). The development of more or less concentric zones representing different stages of contamination resulting from reciprocal reaction between magma and limestone, is illustrated by the diorite intrusion at Ben Bullen, N.S.W., Australia (Joplin, 1935).

Partial melting of country rocks rather than magmatic assimilation has been used to explain the absence, in NE Scotland, of large quantities of transitional types between uncontaminated magmatic rocks and cordierite-bearing noritic rocks that would be expected if the norites were the result of assimilation of pelites (Gribble, 1968). $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the cordierite-bearing xenolithic rocks support a derivation as partial melt products of country rocks without any

contribution from the intruding basic magma (Pankhurst, 1969). Such a process is a likely mechanism of formation of many xenolithic rocks previously interpreted as resulting from contamination of magma by assimilation of host rocks.

The following are other terms used in relation to the process and products of assimilation.

Anomalous—said of a magma type that is the result of assimilation.

Diachyte—a rock produced by marked mechanical and/or chemical contamination of an anatectic magma by cognate basic material.

Endomorphism (endometamorphism)—the process by which an igneous rock is modified due to complete or partial assimilation of country rock fragments or by reaction upon it by the country rock.

Imandrite—a rock composed mainly of quartz and albite formed as the result of assimilation of graywacke by a nepheline syenite magma.

Magmatic dissolution—assimilation of country rock by magma by the process of solution.

Magmasklerosis—a little used term for the magmatic assimilation of limestone.

Syntexis (syntectite)—(1) modification of a magma by assimilation; (2) formation of magma by melting of two or more rock types and assimilation of country rock.

C. D. GRIBBLE

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ASTROBLEME

An *astrobleme* (Greek: star wound) consists of the crater, macro- and microstructures, ejecta and/or shock-metamorphic effects of meteorite impact on the Earth's surface. The energy released by such an event is related to the kinetic energy of the projectile, a function of its mass and terminal velocity ($e = \frac{1}{2}mv^2$, where e = kinetic energy, m = mass, and v = velocity). Thus, a meteorite of radius 10 m (ca. 32,000 t if it is an iron meteorite) traveling at 10 km/sec will release some 1.6×10^{15} joules, approximately the equivalent of the explosion of 380 kt of TNT (about 20 Hiroshima bombs) or an earthquake of 5 on the Richter scale. A crater up to 0.5 km in diameter would be produced if such an impact occurred on land.

In a fall of this size, or larger, the impact would fragment the projectile and pulverize considerable quantities of country rock: small quantities of impact melt material would also be produced. The explosion that excavates the crater, generating the cloud of ejecta, also propagates an intense pressure pulse that penetrates deep into the local bedrock. Four sets of features are produced with quite distinctive characteristics.

1. The crater has a polygonal (and roughly circular) outline. A small crater up to 15 km in diameter will initially have a simple bowl-like profile, although at the larger end of this spectrum the rim wall will rapidly slump inwards owing to slope failure (Fig. 1a). Larger craters (>25 km diameter) are far more complex, often with a large rebound structure in the center and several concentric uplifted rim scarps (Fig. 1b). Craters are relatively superficial features of terrestrial astroblemes, being very susceptible to erosion.

2. The fragmentary material generated by the

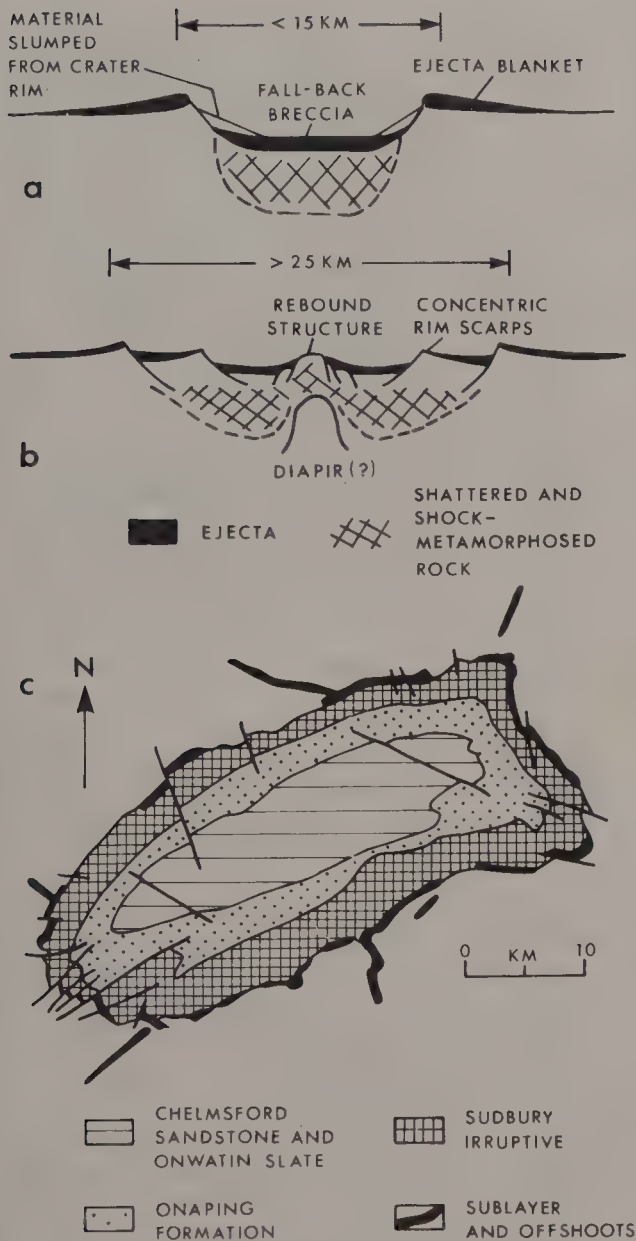


FIGURE 1. (a,b) Schematic sections of small and large impact craters showing distribution of principal features. (c) Map of the Sudbury structure, Ontario, Canada, possibly an astrobleme modified by later tectonism. The Sudbury sublayer includes an impact breccia, shock-metamorphosed material, and a Ni-Cu sulfide-rich rock that may partly be the product of impact melting of basic country rock. The Sudbury Irruptive is a lopolith intruded into the base of the impact basin. The Onaping Formation is a fall-back breccia, and the Chelmsford Sandstone and Onwatin Slate are later sediments deposited in the basin prior to folding and faulting. (After Pattison, 1979.)

explosion (ejecta) consists of both meteoritic material and the excavated and pulverized country rock. Some falls back into the crater (fall-back breccia); the rest blankets the uplifted rim and surrounding area. Most of the ejecta blanket accumulates by airfall, but some outside the crater is deposited by ground-hugging,

high-velocity base surges generated by the blast; the resultant deposits share features in common with volcanic ejecta. Impact ejecta has several unique characteristics: clasts usually show the results of shock metamorphism, while the matrix may consist of either or both comminuted country rock or shock-melted glass. Where the impact affected a layered country rock sequence, the stratigraphy is often preserved, but inverted in the arrangement of clasts in the ejecta. In a number of instances, these distinctive deposits were identified even before their nature was recognized: e.g., *suevite* at Ries Kessel crater, W Germany and *kärnäite* at Lappajärvi, Finland (Fig. 2).

3. *Impactite* (shock-generated glass) consists of both quenched melted rock and material whose crystal structure was totally disrupted by the shock waves, but which was never molten. The former includes *lechatelierite* (pure SiO_2 glass), which can also be produced on a much smaller scale by lightning strikes. Large bodies of impact-melted glass often contain Ni-Fe metal-sulfide spherules with a chemical composition and mineralogy identical to meteoritic material. These are diagnostic and the glass frequently retains crystal shapes, textures and compositions, and is referred to as diaplectic glass. When it is relict, it is termed thetomorphic. *Maskelynite* is a thetomorphic diaplectic glass with the composition of feldspar. Like all *diaplectites* its occurrence is diagnostic of impact.

4. Shock metamorphism produces changes characteristic of very high pressures (usually >30 kb), very high strain rates, and either low or very high temperature (often >1,800°C). Shock metamorphic mineral assemblages are typically not equilibrated. High-pressure polymorphs like coesite and stishovite (SiO_2), or diamond, are all known to occur in other parageneses, e.g., in mantle xenoliths, but their occurrence in shock-metamorphosed sandstone or shale is diagnostic of impact. Very high temperatures (probably as high as 2,000°C), unattainable in magmatic rocks, are indicated by the thermal decomposition of zircon to baddeleyite.

Surface weathering processes, active tectonism and sedimentary reworking of the Earth's surface conspire to destroy the record of terrestrial impact history. Lunar studies, furthermore, indicate that the major phase of bombardment of the terrestrial planets ended some 3-5 Ga ago, so that any record of a similar bombardment of the Earth has probably been erased. Consequently, complete impact craters are rare, and examples like Barringer Crater, Arizona, U.S.A., are relatively young (ca.



FIGURE 2. Worldwide distribution of 90 astroblemes and meteorite impact sites. (Data from Glass, 1982.)

50,000 a old). Generally speaking, the older the astrobleme, the more cryptic the evidence for its impact origin. Examples like Ries Kessel in W Germany, and Bosumtwi in Ghana are less than 15 Ma old but, whilst preserving most of the features outlined above, have outlines that are only conspicuous in aerial photographs. Older examples, such as the 210 Ma Manicouagan structure, the 500–450 Ma Lukanga structure, the 1,840 Ma Sudbury basin and the 1,970 Ma Vredefort dome (Fig. 2), have been more controversial, particularly where igneous activity has occurred. Whether igneous activity can be triggered by massive impact is still contentious.

Over 100 astroblemes have been identified in recent years (Fig. 2 locates 90, including those cited here); their ages range from the 1,970 Ma Vredefort dome to the Sikhote Alin craters, produced by a meteorite fall in 1947. Their distribution partly reflects the extent of detailed aerial surveys, though there is an expected clustering on the older shield areas of Canada and Fennoscandia.

ADRIAN F. PARK

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Cross-references: *Calderas and volcano-tectonic depressions; Cryptovolcanic structure; Explosive volcanic eruptions—classification; Lopolith; Lunar petrology; Meteorites—petrology; Pyroclastic eruptions; Shock metamorphism; Tektite.*

ATOLL GARNET

An *atoll-shaped garnet* usually consists of a complete or almost complete garnet rim enclosing a core that may contain any combination of mica, feldspar, quartz, and iron ore. In some cases double rims occur, i.e., a smaller rim within and separate from the outer rim, and it is also possible to have garnet “islands” within the rims (Fig. 1). Although the

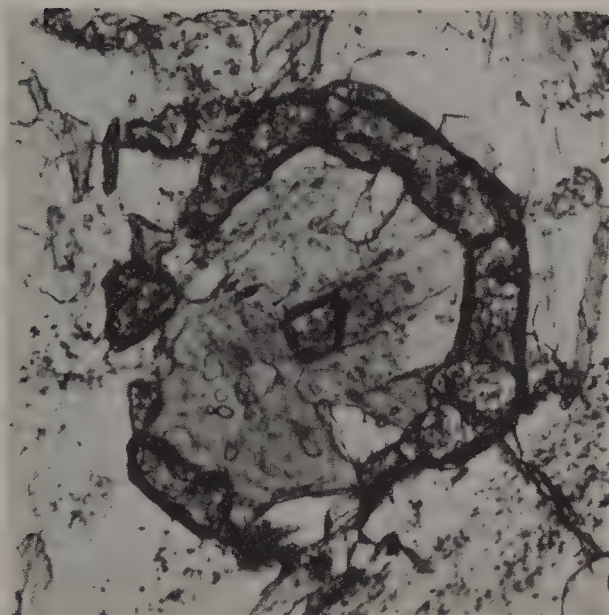


FIGURE 1. Photomicrograph of an atoll-shaped garnet from Ardara, Co. Donegal, Eire; the rim has been breached in several places but the euhedral outline has remained, particularly where in contact with quartz or a quartz-rich matrix; the original garnet has been replaced by biotite plus a little quartz, with the core remaining as a garnet "island"; PPL, $\times 35$.

term "atoll garnet" is descriptive, it is conventional to exclude similar garnet shapes occurring as coronas or reaction rims, or skeletal garnet growths in quartz-rich matrices. However, the growth of a single garnet around grain boundaries may closely resemble an atoll garnet.

Two main origins have been proposed: (1) the rim is older than the core, the latter resulting from the partial replacement of a complete garnet, and (2) the rim is younger than the core, the former resulting from the nucleation and growth of garnet along crystal boundaries. The replacement interpretation is based on textural criteria (Fig. 2; Rast, 1965; Smellie, 1974) with

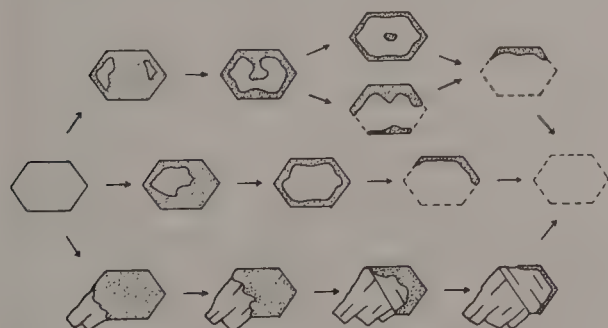


FIGURE 2. Schematic stage-by-stage development of atoll garnets from Ardara; when the garnet finally disappears, a biotite pseudomorph remains that, in some cases, has a euhedral outline, particularly when in a quartz-rich matrix; the speckled areas are garnet that is being progressively replaced.

some atoll garnets having cores of mainly biotite with some quartz, plagioclase, and iron ore, in which crystals of staurolite, andalusite, and sillimanite have grown; this indicates the prograde nature of the garnet core. In contrast, other examples show no evidence of replacement, pointing to garnet growth around single or multiple crystals as a viable explanation.

Chemical evidence may serve as a basis for assessing origin. Provided that a garnet is initially chemically zoned, then subsequent replacement of the core may result in the rim and any residual garnet fragments retaining some of the original compositional characteristics. Microprobe analyses on complete and atoll-shaped garnets led Atherton and Edmunds (1966) and Cooper (1972) to conclude that there was no evidence for replacement and that the atoll rims and garnet "islands" had probably resulted from separate nucleations. In contrast, Smellie (1974) found chemical evidence for replacement that substantiated his textural observations.

The most common occurrence of atoll garnets is in high-grade regional or in thermal metamorphic rocks, indicative of high temperatures of formation.

JOHN A. T. SMELLIE

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Cross-references: *Amphibolite facies*; *Contact (thermal) metamorphism*; *Nucleation of metamorphic minerals*; *Reaction rim*.

AUGITITE

An *augitite* is an undersaturated porphyritic volcanic rock consisting of phenocrysts of augite and iron ore, with or without biotite or hornblende, in a brown glassy (usually Na-rich) groundmass. It was first described by Doelter in 1882 (Holmes, 1928) from the Cape Verde Islands. Hatch et al. (1951) placed augitite in their "fine-grained ultramafite" group formerly

known as “magma basalts,” along with limburgite, olivine melilitite, and related rocks, but did not mention augitite in the 13th edition (1972).

Tröger (1935) gives the mode of augitite from the original Cape Verde Islands as 40% titanaugite, 55% glassy groundmass containing plagioclase and nepheline microlites, with accessory iron ores, apatite, and biotite.

Although probably as close to being a monomineralic lava as any rock, so far as its crystalline part is concerned, augitite is an unfortunate misnomer for a rock consisting of only 40% augite, as it suggests rather a pyroxenite consisting *wholly* of augite.

R. T. PRIDER

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Cross-references: *Limburgite*; *Melilitite*.

AUREOLE

The country rocks adjacent to many igneous intrusions are recrystallized in a zone, known as a *metamorphic aureole*, in which the effects of contact (thermal) metamorphism can be detected; the terms contact aureole and thermal aureole are commonly used. In some aureoles, especially those in argillaceous rocks, concentric zones of progressive metamorphism can be mapped on the basis of mineralogical and textural changes in rocks of the same general composition. Typically, the first development (outer zone) is of ill-defined ovoid spots of graphite, or other minerals (spotted shales and slates), or the growth of porphyroblasts including chiastolite. This is followed by a zone in which extensive recrystallization progressively becomes more apparent and biotite and, in places, andalusite develops. In the inner zone adjacent to granite plutons–batholiths, cordierite is developed in a rock that has become a hornfels (Fig. 1): compact rocks hardened by the action of heat are referred to as having been baked or indurated. Where the temperature close to the contact is even higher, orthoclase and hypersthene occur in the hornfels. Aureoles in calcareous rocks commonly exhibit greater mineralogical variation and zoning is much less well developed; chemical composition, per-

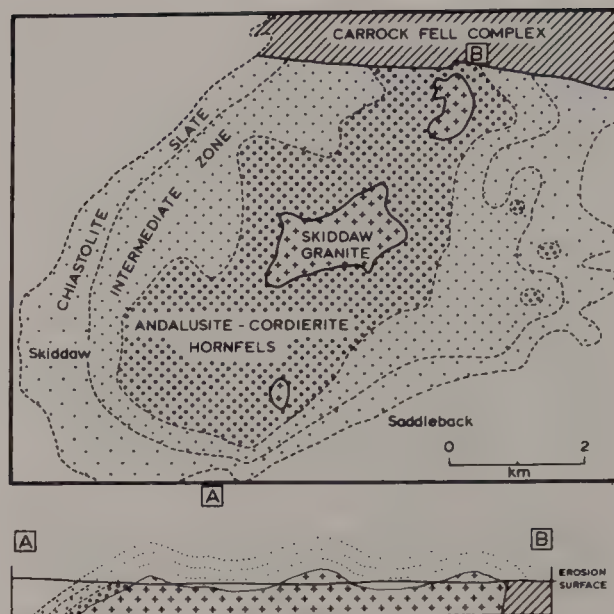


FIGURE 1. Skiddaw granite and its metamorphic aureole, northern England. (After Eastwood, 1953.)

meability of individual calcareous units and variations in the amount of CO_2 present from place to place affect the nature of the mineral assemblage developed. At the contact itself there may be metasomatic introduction into the country rock of certain elements (e.g., Fe, F, B, Cl), leading to the formation of skarns.

The temperature at any distance from the igneous contact is dependent upon the temperature, crystallization temperature, latent heat of crystallization, and size of the body of magma, the thermal conductivity, specific heat, diffusivity, initial temperature, and water content of the country rock, and ΔH of metamorphic reactions set up in the aureole. Aureoles vary from almost insignificant zones of baking next to minor intrusions, such as some sills and dykes, to mineralogically zoned aureoles 1–2 km wide, but showing a much greater outcrop width if the igneous contact is at a low angle. Pore water in the country rocks plays a major role in determining the width of an aureole and the mineral assemblages developed, as heat conducted outwards is used to vaporize pore water as well as to initiate endothermic metamorphic reactions of dehydration and decarbonation. So, while intrusions emplaced into dry rocks commonly show significant aureole effects (Fig. 2a), other intrusions of even higher-temperature magma but emplaced into wet rocks may have very limited aureole effects (Fig. 2b).

Temperatures at contacts with granitic plutons are commonly 500–550°C, with the development of hornblende hornfels facies mineral assemblages in the country rocks. Aureoles are commonly up to 1 km wide with

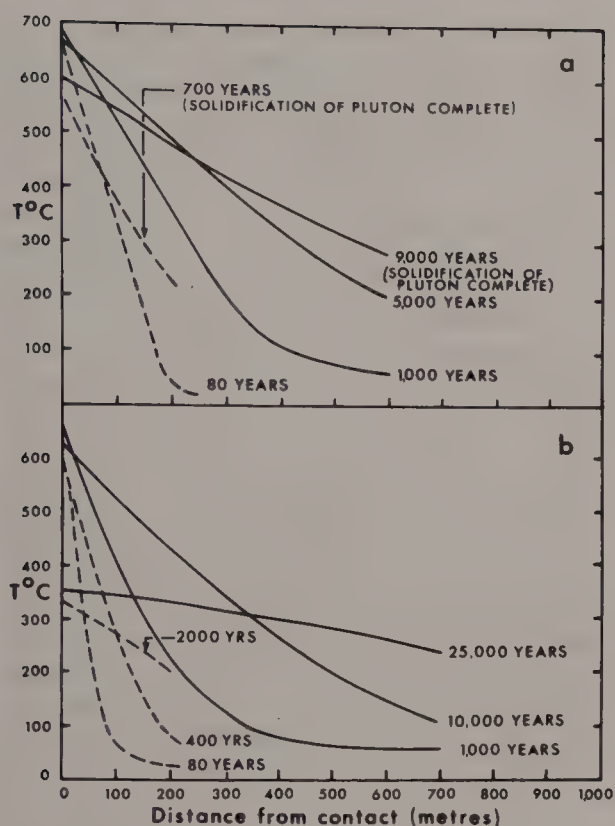


FIGURE 2. (a) Ideal temperature gradients in floor below dolerite sills, neglecting effects of water; intrusion temperature 1050°C, crystallization range in magma 1050–850°C; full lines refer to a sill 700 m thick intruded under a cover of 350 m at 50°C; broken lines refer to a sill 200 m thick intruded under a cover of 100 m at 25°C; times given are in years from date of intrusion. (After Turner, 1968—reference in Turner, 1981.) (b) Ideal temperature gradients adjacent to dolerite sills intruded into water-saturated sediments (porosity 0.2); other conditions as for (a). (After Turner, 1968.)

thermal gradients of ca. 350–400°C/km. Temperatures at contacts with basic plutons under deep cover are commonly 650°–700°C, with the development of pyroxene hornfels facies mineral assemblages near the contact grading into assemblages of lower temperature facies further away in aureoles that are wider than those around granitic plutons. However, many intrusive masses of basic magma are undersaturated in water and emplaced at relatively high levels into water-saturated rocks, while many acidic–intermediate plutons crystallized from magma with a much higher water content which was emplaced, in orogenic zones, into rocks from which much water has been removed during regional metamorphism. Accordingly, while some large basic bodies are surrounded by extensive and spectacular aureoles (e.g., Bushveld Complex, South Africa: Hall, 1932), commonly it is the aureoles around granite–granodiorite–diorite masses that best display the

characteristic zonation (e.g., Comrie, Scotland: Tilley, 1924). Aureoles beneath thick dolerite sills emplaced into water-saturated sediments (e.g., Mt. Wellington sill, Tasmania: Edwards, 1942) are relatively insignificant. In contrast, relatively small minor intrusions of basalt–dolerite that occupy conduits up which there was continued replenishment of hot magma, and particularly of hot gases (e.g., Scawt Hill, Northern Ireland: Tilley, 1929), are surrounded by very high-temperature sanidinite facies mineral assemblages in which the effects of partial fusion may be present (see **Buchite; Pyrometamorphism**). Petrogenetic grids for mineral assemblages below the onset of partial melting and for high-grade assemblages in aureoles are given by Pattison and Harte (1985).

Variations in size and expression of thermal aureoles may reflect subsequent tectonic fragmentation and emplacement of igneous masses and their surrounding rocks (e.g., basic igneous masses in the Caledonides of NE Scotland: Munro and Gallacher, 1984) with some very large ultrabasic masses of “alpine-type” showing little or no aureole effects. Tectonic activity has also been a factor in the development of aureoles beneath obducted slabs of hot mantle material in ophiolite terrane (e.g., Bay of Islands Complex, Newfoundland: see Fig. 3a). Recrystallization of the material over which a hot slab rode resulted in the welding on to its

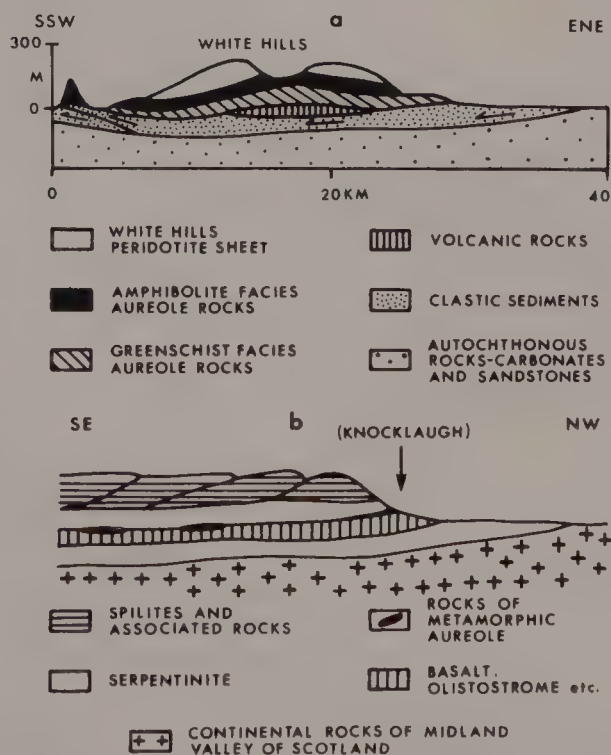


FIGURE 3. Thermal aureoles below obducted ultramafic slabs. (a) Newfoundland. (After Williams and Smyth, 1973); (b) Ballantrae, Scotland. (After Treloar et al., 1980.)

base of successively-formed aureole material representing the changing P - T conditions during upward movement from considerable depths in the mantle. Tectonic telescoping has resulted in the juxtaposition, over thicknesses as little as 50 m, of the products of recrystallization ranging from 1000–900°C to 300–250°C with many of the rocks of the aureole assemblage showing many features in common with mylonites (e.g., Ballantrae Complex, SW Scotland: see Fig. 3b).

Emplacement of magma into tectonically active zones, such as major transcurrent faults, results in some aureole assemblages having mineralogical and textural characteristics similar to those generally associated with regional metamorphism (contact schists); for example, adjacent to much of the Donegal Granite in Eire the aureole consists of schists containing kyanite and staurolite (Fig. 4). Time relations between deformation and recrystallization are shown by differences in texture and mineral composition and a metasomatic zone of fibrolite development overprints the contact metamor-

phic mineral assemblages near the igneous-country rock contact. There, as in other aureoles, features indicative of the mechanism of emplacement of the igneous mass are shown (Pitcher and Read, 1963).

Temperatures and pressures at which recrystallization took place in some aureoles have been calculated using the partitioning of elements in equilibrium mineral assemblages (see **Metamorphic petrology—use of thermodynamics**). For example, garnet-pyroxene and garnet-amphibole Fe-Mg exchange reactions have been used for geothermometry (Treloar et al., 1980) and the assemblage garnet + plagioclase + wollastonite + quartz for geobarometry (Droop and Treloar, 1981). However, in aureoles where the partial pressure of water (and the water content of some minerals) was variable, determination of P - T conditions from place to place on the basis of the development of equilibrium mineral assemblages becomes unreliable (Treloar, 1981). This is certainly the case where meteoric hydrothermal circulation systems played a major role in the

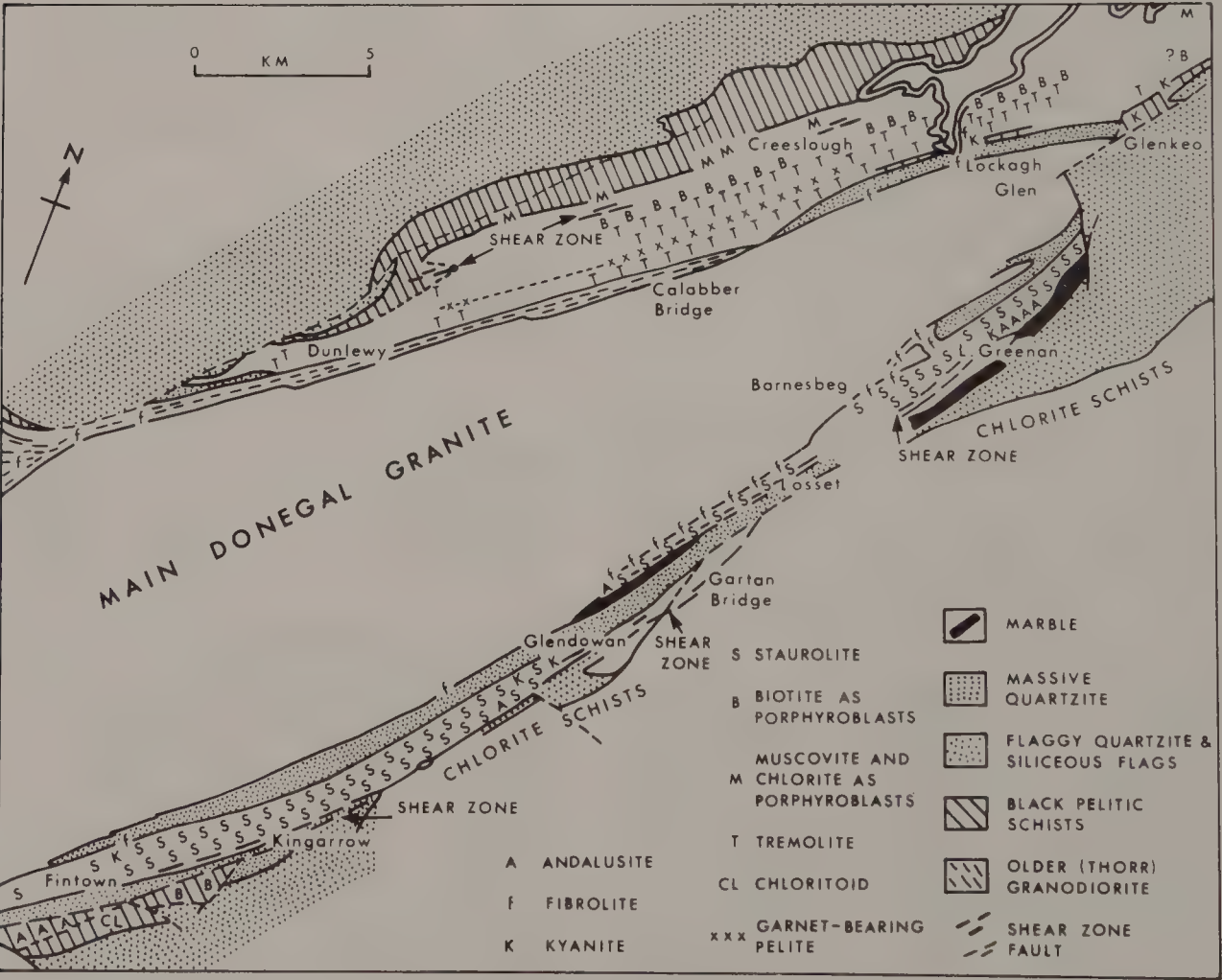


FIGURE 4. Metamorphic aureole of the Main Donegal Granite, Eire. (After Pitcher and Read, 1963.)

cooling of an intrusion. Here, in aureoles adjacent to igneous intrusions emplaced at shallow depths into permeable rocks, there was chemical mass transfer (metasomatism) rather than equilibrium recrystallization. Such hydrothermal convective cooling, which has a marked effect on the maximum temperatures attained at the aureole, may be associated with the transport of ore-forming constituents, for example, in the development of porphyry-copper deposits (Parmentier and Schedl, 1981).

D. R. BOWES

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Cross-references: *Batholith*; *Buchite*; *Contact (thermal) metamorphism*; *Contact metasomatism*; *Facies of thermal metamorphism*; *Hornfels*; *Hydrothermal processes*; *Metamorphic mineral growth*; *Metamorphic petrology—use of thermodynamics*; *Ophiolite*; *Pyrometamorphism*; *Skarn*.

B

BANDING IN IGNEOUS ROCKS

The common everyday usage of the term “banding” would suggest that it should be employed to describe essentially two-dimensional, strip-like features of igneous rocks. However, it has come to be used by geologists to describe igneous rocks that contain macroscopic inhomogeneities with an essentially sheet-like form, although it should be noted that in many outcrops of limited size these inhomogeneities may appear to be two-dimensional. The inhomogeneities always have a measurable thickness (normally ranging from a few millimeters to several metres) although this dimension is generally smaller by at least an order of magnitude than the length and breadth of the “band.” Structures, such as foliation, that are defined simply by the orientation of crystals and inclusions are thus excluded from this discussion. “Bands” may show varying degrees of regularity and persistence: in some cases they are formed by extensive planar features of constant thickness, but in others they are defined by inhomogeneities that are markedly irregular and impersistent and are often folded and convoluted. Planar features produced by deformation unrelated in origin to the igneous rock (e.g., shear zones produced during later deformation) are not considered here.

Two main types of inhomogeneity can be recognized, viz., structural and mineralogical (lithological).

Structural banding

Rocks with the type of inhomogeneity known as *structural banding* display variations in structure and texture that are not accompanied by appreciable variations in mineralogy and chemistry. Such banding is well developed in many glassy or fine-grained siliceous rocks that tend to occur in shallow intrusions or as extrusions (e.g., rhyolites—see **Endogenous dome**). In these rocks the banding is due to variations in color and texture, with adjoining layers showing marked variations in crystallinity (some being glassy, others relatively well crystallized), in content of vesicles, or in the degree to which spherulitic growths or lithophysae have developed. These bands may

be relatively regular in form or they may be highly convoluted (Fig. 1). The origin of this type of banding is not properly understood, but it has been suggested (e.g., Macdonald, 1972) that it is produced as the result of the deformation of a body of viscous magma within which there are considerable variations in volatile content. In some instances the attitude of the banding shows similarities to that of the structures in valley glaciers. Structures of this type can be recognized not only in glassy rocks that appear to have quenched directly from magma, but also in ignimbrites. However, the same explanation, involving deformation due to viscous flow, is generally invoked in both cases to account for the banding.

Mineralogical banding

Mineralogical banding is varied in character and in origin.

Cumulate layering. Wager and Brown (1968) discuss the use of the term “banding” and “layering” and suggest that *banding* should be used for any streaky or roughly planar heterogeneity in igneous rocks, the term *layering* being reserved for sheet-like structures produced by crystal settling. Cumulate layers show many similarities to sedimentary structures with segregation of different crystal phases according to size and density, often yielding monomineralic layers or producing structures resembling graded bedding. Support for the crystal-settling hypothesis is also provided by the unique textural features displayed by cumulates (see **Cumulate textures**; **Intrusion—layered**).

Willow Lake-type banding. Taubeneck and Poldervaart (1960) described a small basic intrusion in Oregon, U.S.A., that not only contains cumulate layering, but also displays banding of a different type that does not seem to involve crystal settling and accumulation. This structure is defined by layers of differing mineralogy in which the crystals (particularly of plagioclase) tend to be elongate and to be oriented perpendicular to the surface of the unit (Fig. 2). Some are very thin and effectively with the thickness of a single crystal; others, which tend to be rich in pyroxene or hornblende, are much thicker. The bands are often regular and

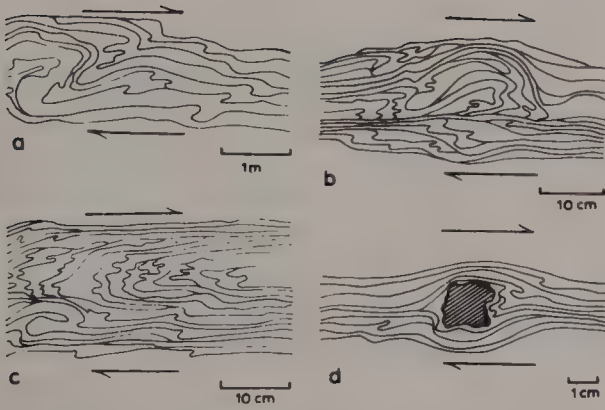


FIGURE 1. Convoluted flow banding in rhyolite lava flows. (After Christiansen and Lipman, 1966.)

parallel, but in some places they are folded or injected and brecciated by normal unbanded norite or gabbro. Bands of similar type are also found enclosing xenoliths and in some cases a fragment of banded rocks is encircled by later banding. These features suggest that this type of banding may be related in origin to some of the orbicular structures found in igneous rocks. The rhythmic banding is interpreted as being due to a cyclic repetition of local undercooling, either at the margins of the mass or near a xenolith, coupled with the effect of intermittent turbulent and convective currents in the magma. A similar mechanism has also been invoked to account for the growth of very elongate crystals (crescumlites) perpendicular to cumulate layers or at the margins of bodies of cumulates (Wager and Brown, 1968) during periods when the magma is thought to be relatively immobile.

Banded minor intrusions. A number of relatively small basic and ultrabasic intrusions, mainly sills and dykes, display internal lithological variations that give these bodies a rather ill-defined banded character (Figs. 3, 4). These "bands," which are generally due to portions of

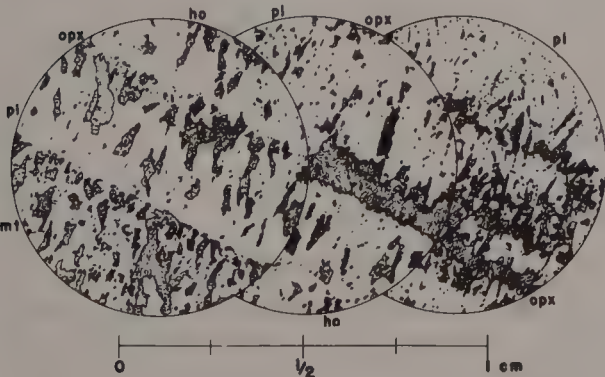


FIGURE 2. Willow Lake-type banding: pl, plagioclase; opx, orthopyroxene; mt, magnetite; ho, hornblende. (From Taubeneck and Poldervaart, 1960.)

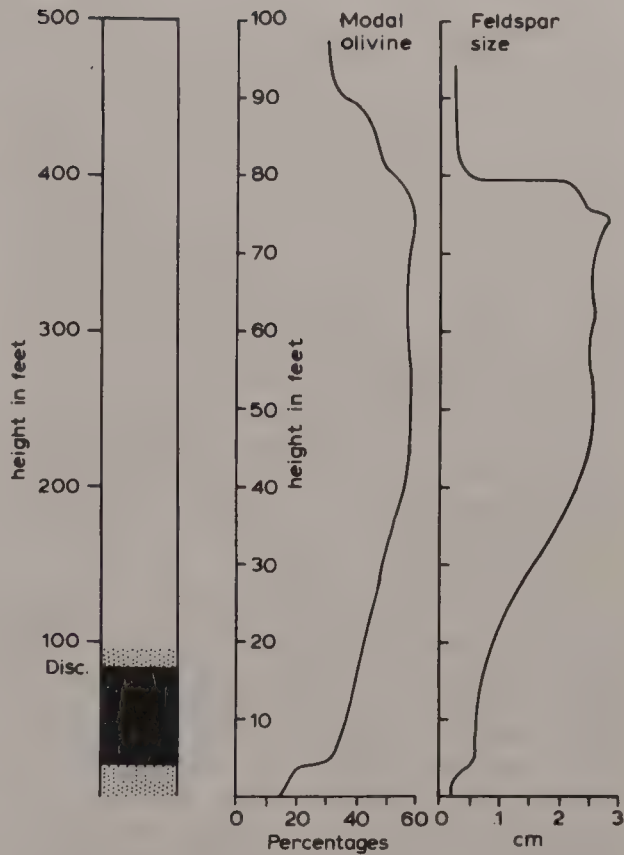


FIGURE 3. Cross-section of the Shiant Isles alkaline dolerite sill with picritic (black) and transitional (stippled) facies: Disc., discontinuity. (After Drever and Johnston, in Wyllie, 1967.)

the intrusion being rich in olivine, tend to be parallel to the general trend of the margins of the intrusion and in undisturbed intrusions are thus approximately horizontal in sills and approximately vertical in dykes. In certain sills it seems that the presence of a relatively thin



FIGURE 4. Flowage differentiation in model dykes and sills containing solid-fluid mixtures simulating olivine phenocrysts and basic melt; solid particles appear black; feeder is 2.5 cm wide. (a) 50% by volume of solid, flow rate 4 mm/sec; (b) 15% by volume of solid, flow rate 3.5 mm/sec, dropping to 1.8 mm/sec. (From Bhattacharji, 1967.)

olivine-rich band (layer) near the base of the intrusion is due to crystal accumulation, notably in the Palisade sill, New Jersey, U.S.A. However, it is difficult to visualize how the banding in dykes could be produced by an accumulation process. In addition, in many sills, particularly of alkaline basic rocks, the olivine-rich band or bands lie at relatively high levels in the intrusion or form too large a proportion of the intrusion for straightforward crystal settling from basic magma to be envisaged (Fig. 3). An alternative explanation is that this banding was produced by a process termed *flowage differentiation*. This process, which has been simulated in laboratory experiments under certain conditions, leads to the central concentration of solid particles held in suspension in a liquid being intruded into a narrow vertical fissure (Fig. 4). The effect is due to the strong velocity gradients near the margins of the intrusion, which cause the suspended particles to be rotated and to migrate towards the center of the body. Continued intrusion as a sill will sometimes superimpose the effects due to gravity on that due to flowage differentiation, leading to the portion of magma enriched in early-formed crystals (e.g., olivine) being displaced towards the lower part of the intrusion (Figs. 3, 5).

Banded granitic rocks. A number of granitic intrusions display banding (not foliation) that may have a variety of origins. In some instances it appears to be a relict country rock structure implying extensive metasomatism and replacement and is thus not truly igneous in origin; in others it seems to be due to the streaking out of aggregates of early-formed crystals (schlieren). Another type has been described from Donegal, Eire, where distinct bands of finer-

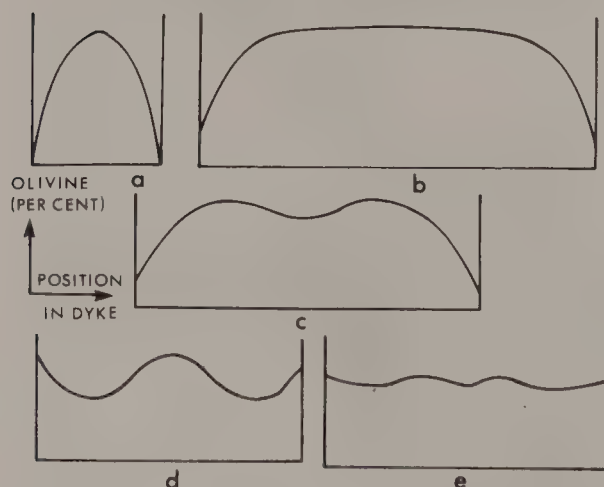


FIGURE 5. Variations in olivine content in five dykes, Skye, Scotland; olivine contents vary from 0 to 70%; dyke widths up to 10 m. (After Gibb, 1968.)

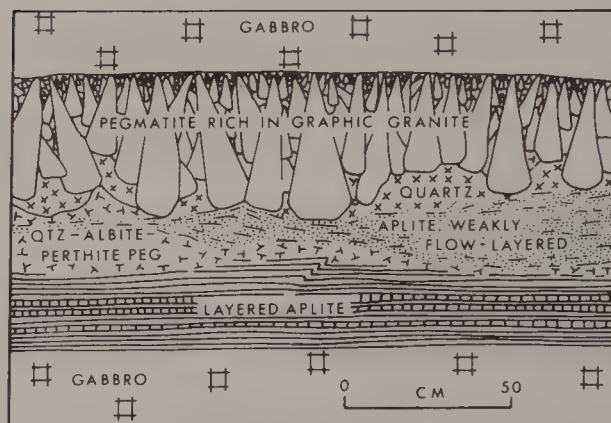


FIGURE 6. Asymmetrical pegmatite-aplite intrusion, Mesa Grande district, California. (After Jahns and Tuttle, 1963.)

grained dark rock that are poor in K-feldspar are associated with, and are sometimes cut by, bands of coarser-grained, paler rock that are rich in K-feldspar (Pitcher and Berger, 1972). It is suggested that these bands were produced by the addition of K-rich fluids to aggregates of crystals of quartz, plagioclase, and biotite (the dark bands) along planar zones of movement.

A similar process of diffusion of late K-rich fluids has been invoked to account for certain banded pegmatite-aplite bodies that show marked asymmetry: the upper part consists of perthite-rich pegmatite and the lower part of albite-rich aplite (Fig. 6). Other banded pegmatite-aplite bodies appear to be due to multiple intrusion of residual granitic magmas of varying character, while still others appear to have acquired a banded character following a single intrusive event. In these bodies, layers rich in minerals such as tourmaline and mica are thought to have formed as the result of periodic supersaturation (cf., Willow Lake-type banding), while interlayered units of pegmatite and aplite are thought to have formed as the result of variations in volatiles content (and confining pressure). The coarse-grained pegmatite forms from volatile-rich magma at high pressure, while the finer-grained aplite forms when volatile loss following a pressure drop leads to rapid crystallization.

M. MUNRO

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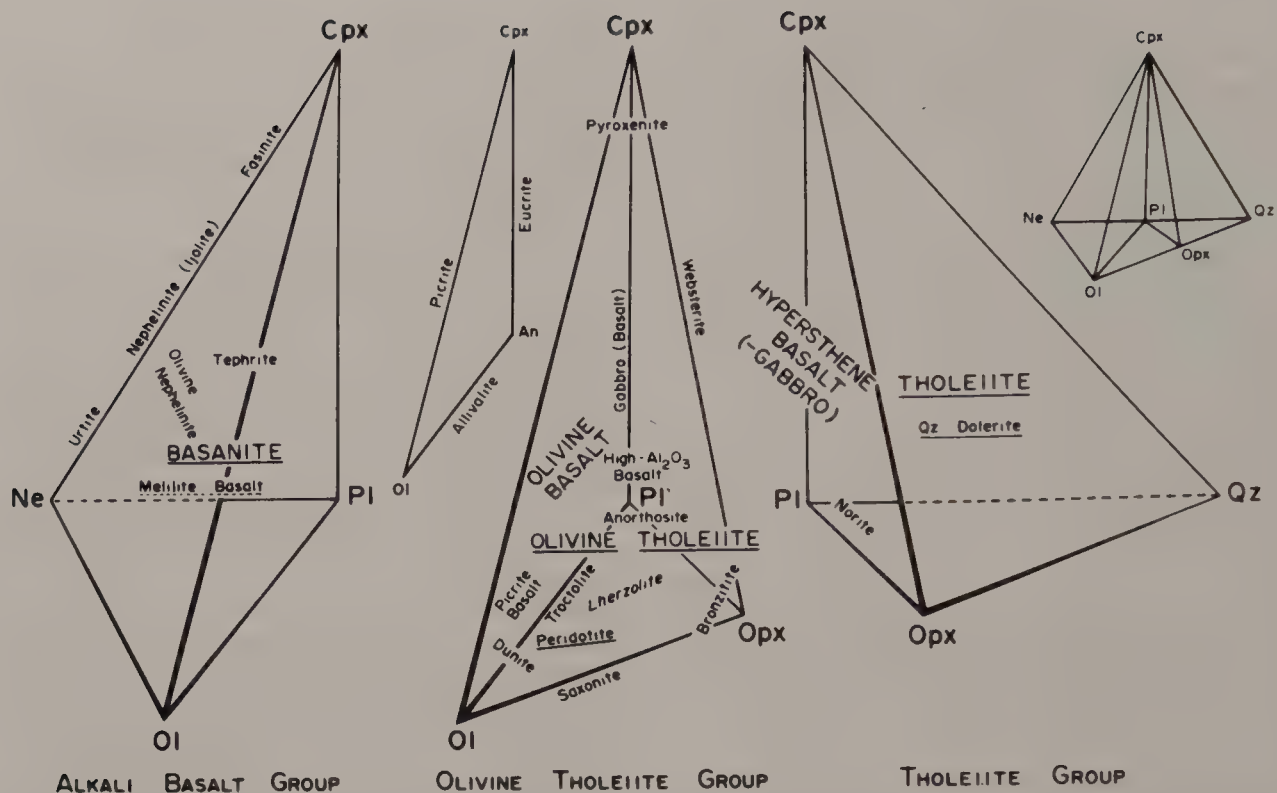


FIGURE 2. Exploded view of generalized simple basalt system (see inset) with the names of rocks whose major normative phases are contained in the tetrahedra; names underlined are within the tetrahedra, names on the faces are written parallel to the base. Lherzolite is in the Ol-Cpx-Opx face. An additional inset, Ol-Cpx-An, gives an alternative nomenclature for when plagioclase is rich in An. Cpx, clinopyroxene; Opx, orthopyroxene; Ol, olivine-forsterite; Ne, nepheline; Pl, plagioclase; An, anorthite. (From Yoder and Tilley, 1962.)

erupted at mid-ocean ridges and underlie much of the ocean floor. Variants include high-Al basalts and high-Mg (picritic) basalts. Further subdivision includes “normal” or N-MORB types found on most ridges, “plume” or “anomalous” P-MORB types found on mid-ocean ridge “hot-spots” such as Iceland, the Azores and Galapagos islands and “transitional” or T-MORB types.

2. *Island-arc basalts* (IAB)—predominantly tholeiites, but carrying more variable K contents and Al abundances; these are sometimes subdivided into low-K tholeiites, high-Al basalts, and high-K basalts or shoshonites. The latter two are often included in the calc-alkaline suite and referred to as calc-alkaline basalts.
3. *Ocean-island basalts* (OIB) include tholeiites and alkaline and olivine-bearing basalts, e.g., Hawaii, Canary Islands, Reunion. Though tholeiites are present, their evolution generally follows alkaline trends.
4. *Continental plateau basalts* (CPB) are represented by such provinces as the Columbia and Snake River basalts of NW United States. Tholeiites may predominate, and evolution may follow a tholeiitic trend, but

alkaline basalts and alkaline trends are not uncommon.

OIB and CPB both occur over mantle plumes or “hot spots” and are sometimes grouped as “within plate basalts” (WPB). Basalts are also erupted in “back-arc” or “marginal” basin environments, e.g., Sea of Japan, Scotia Sea. Their properties are intermediate between MORB and IAB (cf., Weaver et al., 1979).

Petrography and mineralogy

All basalts contain calcic plagioclase (An_{50+}), usually labradorite, existing as laths, often defining an ophitic or subophitic texture. The feldspar laths are enclosed in chadocrysts of clinopyroxene. Other phases include a second pyroxene, Fe-Ti oxides and apatite. Glass is a common interstitial component with variable composition.

Tholeiite. The groundmass consists of plagioclase laths embedded in clinopyroxene, usually augite. Olivine may be present as phenocrysts or xenocrysts, but it is not in equilibrium with the groundmass, and usually carries a reaction rim of Ca-poor pyroxene (enstatite, bronzite, or subcalcic augite). Interstitial glass may be

present, with a high SiO_2 content, though often this is replaced by quartz.

Hypersthene basalt. This differs from tholeiite only in its pyroxene composition. Olivine phenocrysts are rare and quartz is usually absent.

Olivine tholeiite. The groundmass contains olivine and pyroxenes: both augite and hypersthene may be present, as well as pigeonite, which may be replaced by complex intergrowths of augite, diopside, and clinoenstatite. Quartz is usually absent.

Olivine basalt and alkaline basalt. Olivine occurs both as phenocrysts and a groundmass phase. The pyroxene is a diagnostic titanite (Ti-rich), to the exclusion of hypersthene and pigeonite. Interstitial phases include alkali feldspar, nepheline, leucite, and analcime. Quartz is invariably absent.

With decreasing An content in plagioclase and the appearance of alkali feldspars, tholeiite, hypersthene basalt, and olivine tholeiite grade into andesite. Alkaline basalt likewise grades into trachyte and phonolite, while with increasing feldspathoid content it grades through basanite to tephrite. Increasing modal mafic minerals produce picrite (olivine-rich basalt) and ankaramite (clinopyroxene-rich basalt).

Lunar basalts. These display some mineralogical, chemical, and textural similarities to terrestrial basalts. The essential minerals are olivine, clinopyroxene (including pigeonite, and all are more Fe-rich than in terrestrial basalts), plagioclase (An_{75-96} , often as phenocrysts), and ilmenite. Magnetite is replaced by more Al- and Cr-rich spinels. Unique lunar phases include pyroxferite (a pyroxenoid), armalcolite and whitlockite. No hydrous or oxidized phases have been identified. Lunar basalts are either olivine-, hypersthene-, or quartz-normative.

Chemical composition

Major and trace element data on selected basalt types are listed in Table 1. Distribution of Fe, Mg, and Na + K (alkalies) is usually illustrated on an A (alkalies)—F($\text{Fe}_2\text{O}_3 + \text{FeO}$)—M(MgO) diagram. Subalkaline and alkaline basalts show wide variation in Na/K, Fe/Mg ratios and Al_2O_3 content. Tholeiitic basalts show the characteristic relationship between SiO_2 content and Fe/Mg ratio, with high- SiO_2 types being markedly Fe-enriched. Independently, tholeiites show a large range in Al_2O_3 content (from high-Al with 15–19 wt.% Al_2O_3 to low-Al with ca. 12 wt.% Al_2O_3). Likewise the Na/K ratio varies considerably, up to 10, but the actual K_2O content and total alkali content are not independent of SiO_2 content, or Fe/Mg ratio. Shoshonites, for

instance, are typically high-Al, high- K_2O , high- SiO_2 rocks and have high Fe/Mg ratios, whereas ocean-floor tholeiites are low-Al, low- K_2O rocks with low Fe/Mg ratios. Alkaline basalts have a higher total alkali content than other basalts with the same SiO_2 content. Generally, Na predominates over K, though there are significant exceptions, like the high-K basalts of the Mediterranean area. High-Al alkaline basalts are rare; the Ca content falls at high-Al levels, and high-Al alkaline basalts have their equivalent in Fe-poor trachybasalts.

Trace-element chemistry does not generally reflect the petrographic subdivision used above, but significantly reflects tectonic setting. MORB is considered the most primitive basalt with respect to a chondrite standard. It is characterized by high Cr and Ni contents, but low abundances of the incompatible elements, such as Rb, Ba, Cs, Th, U, P, Sr, Pb, Zr, and Nb, compared to other basalts. K/Rb ratios exceed 1000, and rare-earth elements (REE) show no concentration. Chondrite normalized REE patterns are typically flat at low abundances being only slightly higher than chondrite itself. MORB has low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (<0.703) and high $^{143}\text{Nd}/^{144}\text{Nd}$ ratios. Anomalous P-MORB has high concentrations of incompatible elements and high light REE, with more radiogenic Sr, Pb, and Nd.

IAB have more complex chemistry, similar in their trends to those of calc-alkaline rocks. Compared to MORB they are enriched in the large-ion lithophile (LIL) elements (K, Rb, Ba, Cs, Sr) relative to the high-field-strength elements (HFSE) (Ti, Zr, Y) and the REE. Chondrite normalized REE patterns are either flat and similar to MORB, or show some light REE enrichment. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios cover the same range as MORB, but have a higher upper limit.

OIB include both tholeiites and alkaline groups. The tholeiites show many similarities to MORB, but are more enriched in LIL, some HFSE and have higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The isotopic ratios of Sr, Pb, and Nd differ considerably between individual island centers.

CPB, both tholeiitic and alkaline, have different chemistries from the OIB equivalents. They have high SiO_2 , K_2O , and Na_2O contents, and higher K/Na and Fe/Mg ratios. There is greater enrichment in the incompatible elements Rb, Sr, Ba, Th, U, and REE, and though the K/Rb ratio in alkaline CPB is comparable to alkaline OIB, it is considerably lower in tholeiitic CPB than tholeiitic OIB. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are higher in CPB than OIB, whereas the $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are lower: this has been taken to indicate contamination of the parent magma by continental crust material.

TABLE 1. Chemical composition of basalts

	Major elements (wt.%)						
	1	2	3	4	5	6	
SiO ₂	49.34	49.36	49.50	45.40	50.62	52.12	
TiO ₂	1.49	2.50	1.60	3.00	0.83	1.07	
Al ₂ O ₃	17.04	13.94	15.50	14.70	16.01	18.07	
Fe ₂ O ₃	1.99	3.03	3.80	4.10	4.11	—	
FeO	6.82	8.53	7.80	9.20	4.55	8.89 ^a	
MnO	0.17	0.16	0.20	0.20	0.17	—	
MgO	7.19	8.44	6.20	7.80	6.24	5.00	
CaO	11.72	10.30	10.20	10.50	9.26	9.67	
Na ₂ O	2.73	2.13	2.80	3.00	2.93	2.95	
K ₂ O	0.16	0.38	0.60	1.00	2.74	0.94	
P ₂ O ₅	0.16	0.26	0.30	0.40	0.44	—	
H ₂ O	1.27	—	1.60	—	1.62	—	
CO ₂	—	—	—	—	0.16	—	
TOTAL	100.08	99.03	100.10	99.30	99.68	98.71	
	Trace elements (p.p.m.)						
	7	8	9	10	11	12	13
Sc	—	30	—	20	25	35	35.5
V	—	251	—	213	290	320	196
Cr	297	168	79	202	156	57	378
Co	32	48	—	43	24	31	—
Ni	97	134	44	145	50	23	221
Cu	—	90	—	85	159	111	24
Rb	<10	22	11	32	59	32	7
Sr	130	328	241	702	943	400	444
Y	43	28	—	33	16	28	18
Zr	95	137	160	189	67	129	92
Nb	—	13	—	69	—	5.8	<4
Cs	—	1.1	—	—	0.85	1.5	<0.6
Ba	14	246	440	528	683	340	161
Ce	—	33	—	105	—	—	22
Th	—	—	—	—	1.28	3.6	0.44
U	—	—	—	—	0.57	0.75	0.33

^a Total Fe as FeO.

1. Average ocean tholeiite (Engel et al., 1965).
2. Average Hawaiian tholeiite (Macdonald and Katsura, 1964).
3. Continental flood basalt; average Picture Gorge type, Columbia River Plateau, NW U.S.A. (Waters, 1961).
4. Average of 35 Hawaiian alkaline olivine basalts (Macdonald, 1968).
5. Average of 42 shoshonitic basalts (Morrison et al., 1980).
6. Average of 75 calc-alkaline basalts (Pearce, 1976).
7. Average ocean tholeiite (Engel et al., 1965).
8. Average tholeiite (Wedepohl, 1975).
9. Continental flood basalt; average Picture Gorge type, Columbia River Plateau, NW U.S.A. (McDougall, 1976).
10. Average alkaline olivine basalt (Wedepohl, 1975).
11. Island-arc shoshonitic basalt (average of 33) (Morrison et al., 1980).
12. Average calc-alkaline basalt, Java and Bali (Whitford et al., 1979).
13. High-Al basalt, Mexico (Luhr and Carmichael, 1981).

Petrogenesis

Basaltic magmas are most probably generated by the partial melting of peridotite at mantle depths (Bowen, 1928; Green, in Hess and Poldervaart, 1967: for comprehensive reviews and extensive references, see Kushiro, 1979,

and Morse, 1980). Experimental petrology has contributed much to fundamental understanding of the operation of this process, with experiments carried out on the melting of basalt at 1 atm pressure, on natural peridotite at 1 atm, on natural peridotite and synthetic pyrolite under a range of pressures and dry conditions,

or under controlled pressures for H_2O and other volatiles.

The dry solidus temperature of peridotite is very high, and increases with depth, so that it is unlikely to be reached along an undisturbed geothermal gradient in any geotectonic setting. Various mechanisms have been proposed to resolve this paradox, of which two are consistent with geophysical, chemical, and isotopic lines of evidence. A mantle diapir, solid, and rising under adiabatic conditions, may permit peridotite to exceed the solidus, and adiabatic cooling could provide the heat for melting. Such melting lowers the density of the diapir, promoting further partial melting. The rise only ceases when the melt fraction segregates and separates. Such a mechanism is consistent with the concepts of mantle plumes beneath "hot spots," and mantle upwellings beneath mid-ocean ridges. An influx of volatiles may similarly depress the peridotite solidus, with 0.1% H_2O significantly lowering the solidus for garnet lherzolite. Such a mechanism may be important in the mantle wedge above subducted ocean crust, which, while undergoing metamorphism under mantle conditions, dehydrates and liberates water. CO_2 may have a similar effect at depths below 75 km.

Physical and chemical conditions of initial melting. Pressure and volatile partial pressure are the dominant factors in determining the position of the peridotite solidus. Lithostatic pressure determines the phase composition of mantle peridotite, so that discrete depth facies are recognized; with increasing depth these are plagioclase peridotite, spinel lherzolite and garnet lherzolite. Melting of mantle peridotites is incongruent, insofar as the resultant liquid and residue are not the same in composition as the parent, though some phases melt congruently. Thus the composition of near-solidus liquids becomes more olivine-rich with increasing depth, and less aluminous. Plagioclase peridotite melting at 5–10 kb may provide the parental liquid for high-Al basalts, while deep melting of garnet lherzolite may yield picrite.

Volatiles can depress the solidus temperature, and in this respect H_2O , F, and CO_2 are considered important. Their physical state is also crucial, namely, whether they are present as an intergranular "pore" fluid, or bound into volatile-bearing phases such as phlogopite, amphibole, or carbonate. The presence of such phases can radically affect the partitioning of certain elements, such as K, Al, Ti, REE, and radiogenic elements, between solid and melt during incongruent melting.

The degree of partial melting of standard garnet lherzolite or pyrolite affects melt

composition. Elements such as K, Na, Al, Ca, Fe, and Ti, and the LIL elements partition readily into the melt fraction, whereas Si, Mg, Cr, and Ni behave in the opposite way. Thus, at low degrees of partial melting the liquid will be silica-undersaturated, strongly alkaline, and ultrapotassic, and enriched in feldspar and incompatible elements. As the degree of melting increases, the liquid becomes more silicic, until, for fractions greater than 20% when phases like garnet and clinopyroxene are completely eliminated from the residue, the melts approximate to hypersthene-normative or tholeiitic liquids. In general, it appears that nepheline-normative basaltic magmas are produced at greater depths and involve smaller degrees of partial melting than hypersthene-normative liquids, which are, in turn, generated deeper than quartz-normative liquids (see Kushiro, 1979; Morse, 1980).

Subsequent relationship between partial melt and mantle material. In a rising mantle diapir the buoyancy is augmented by partial melting. When the melt initially forms it is in equilibrium with the solid residue, but this status is pressure-dependent. Unless the melt segregates and separates from the residue, continued rise of the total mass into lower-pressure regions permits the modification of the melt, either by added melt increments from the further fusion of solid mantle residue, or re-equilibration between melt and residue. As an example, the formation of komatiite liquid may involve the dissolution of a harzburgitic residue into a tholeiitic liquid.

History of the melt after segregation and separation. Processes concerning the behavior of the basaltic melt and the nature of its interactions with the medium through which it migrates influence the nature of the liquid. These are as follows.

Crystal fractionation and other differentiation processes affect magma in large bodies. These are diverse, but their exact nature can produce a series of distinct trends in rock composition derived from a relatively small number of parent compositions (see **Magmatic differentiation, Calc-alkaline rocks, Feldspathoidal rocks—genesis**).

Assimilation of material from the rocks through which the magma migrates will alter its composition. The scale of this process is constrained by the heat-consuming nature of this process, but crustal contamination is held responsible for some chemical and isotopic characteristics of erupted basalts in the island arc and continental plateau environments.

Hybridization and crustal anatexis involve a rising basaltic magma providing the heat source for the melting of significant quantities of

continental crust, and the mixing of this silicic melt with the basalt liquid to varying degrees. Again, thermal constraints limit the scale of this process, but it is regarded as important in certain provinces (see **British Tertiary Volcanic Province**).

Effects of mantle source heterogeneity

Petrogenetic models that assume an homogeneous mantle source, while useful, are no doubt oversimplified. The study of mantle xenoliths and magmas derived directly from the mantle has revealed considerable heterogeneity (see reviews in Hawkesworth and Norry, 1983). Recent models consider a layered mantle with an upper layer from which basalts have been removed (regarded as depleted compared with primordial or chondritic compositions) and a lower layer closer in composition to the chondritic standard. Thus, basalts derived from the upper mantle—e.g., MORB and IAB—have trace-element, REE, and isotope signatures reflecting a depleted source. OIB and CPB, on the other hand, carry evidence of a selective enrichment in certain elements, notably light REE and LIL elements. At one time this was regarded as a consequence of the physical mixing of undepleted and depleted mantle material in a rising plume. More recent models, derived from the study of xenoliths in kimberlite pipes, prefer a vein-borne metasomatism of the upper mantle by LIL- and light-REE-enriched fluids originating deeper down (see Menzies, in Hawkesworth and Norry, 1983).

Other terms

Absarokite—a basaltic rock, composed of olivine and augite phenocrysts in a groundmass of labradorite with orthoclase rims, olivine, and some leucite; it may grade into shoshonite and into banakite (Absaroka Range, Wyoming, U.S.A.).

Alboranite—an olivine-free hypersthene-bearing basalt (Alboran Is., Spain).

Anamesite—a basaltic rock intermediate between aphanitic basalt and dolerite.

Arapaphite—a magnetite (ca. 50%)—bytownite–augite basalt.

Augitophyre—a porphyritic basalt with augite phenocrysts.

Banakite—composed of olivine and augite phenocrysts in a groundmass of labradorite with orthoclase rims, olivine, augite, and leucite; grades into shoshonite and absarokite (Bannock Indians, Yellowstone Park region, U.S.A.).

Basaltite—olivine-free basalt.

Basanoid—(1) an alkali basalt free from nepheline but with a Na-rich glassy ground-

mass; (2) a basanite-like rock having normative but no modal feldspathoids.

Batukite—a melanocratic leucite basalt; augite and olivine phenocrysts in a groundmass of augite, magnetite, and leucite (Batuku, Celebes, Indonesia).

Carmeloite—a basalt or andesite (depending on whether the plagioclase is labradorite or andesine) with iddingsite presumed to be after olivine (Carmel Bay, California, U.S.A.).

Etnaite—an alkali olivine basalt.

Ghizite—an analcite basalt with biotite.

Hawaiite—an andesine-bearing alkali olivine basalt; intermediate between alkali basalt and mugearite (Hawaiian Is.).

Linosaite—a basaltic rock with alkaline affinities containing Na-pyribole or a minor proportion of feldspathoid (Linosa, Mediterranean Sea).

Macedonite—an aphanitic basaltic rock with feldspars, nosean, melilite, perovskite, rare pyriboles, serpentine or chlorite as pseudomorphs after olivine, and a green vitreous or chloritic base (Mt. Macedon, Victoria, Australia).

Miharaite—an alboranite with quartz in the groundmass; a quartz-bearing hypersthene basalt (Miharayama, Japan).

Mijkite—a Mn-rich basalt with augite and bytownite phenocrysts in a groundmass, with intersertal texture, composed of feldspar, magnetite, red-brown Mn-pyroxene, and glass (Mijkakeshima, Japan).

Onkilonite—a nepheline–leucite basalt with olivine, augite, perovskite, and interstitial glass, but no feldspars or Fe-ores.

Nonesite—a porphyritic basalt with enstatite, labradorite, and augite phenocrysts in a plagioclase–augite groundmass.

Palatinite—a little-used term for orthopyroxene-bearing basalts and andesites.

Picrite basalt—a basalt rich in olivine; *oceanite* if phenocrysts mainly olivine; *ankaramite* if phenocrysts mainly olivine and augite.

Scanoite—a basaltic rock similar to ghizite but containing normative nepheline.

Schönfelsite—a very dark-colored bytownite (ca. 28%)–olivine–augite–bronzite basalt (Altschönfels, W Germany).

Shoshonite—a basaltic–andesitic rock with olivine and augite phenocrysts in a groundmass of labradorite with orthoclase rims, olivine, augite, and small proportions of leucite and dark-colored glass; grades into absarokite and banakite (Shoshone River, Wyoming, U.S.A.).

Sudburite—an augite–hypersthene basalt with pillow structure (Sudbury, Ontario, Canada).

Tannbuschite—a dark-colored nepheline basalt.

Vicoite—a leucite-bearing shoshonite containing leucite, calcic plagioclase, sodic sanidine, and augite (Vico Volcano, Italy).

Weiselbergite—an altered glassy basalt with phenocrysts of labradorite, augite, and opaque ores in a groundmass of plagioclase, augite microlites and interstitial glass.

Woodenite—a basaltic rock containing olivine and augite in an alkalic, brown, glassy groundmass and a chemical composition similar to that of absarokite (Woodend, Victoria, Australia).

Yamaskite—(1) a basalt containing hornblende, titanite, a small proportion of anorthite, and accessory biotite and Fe-ores; (2) an amphibole-bearing jacupirangite (Mt. Yamaska, Quebec, Canada).

E. JÉLINEK
F. V. HOLUB
H. KLÁPOVÁ
J. SOUČEK

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Cross-references: *British Tertiary Volcanic Province*; *Calc-alkaline rocks*; *Feldspathoidal rocks—genesis*; *Flood basalt*; *Igneous rocks—tectonic setting*; *Island arcs—petrology*; *Lunar petrology*; *Mafic and ultramafic inclusions in igneous rocks*; *Magmatic differentiation*; *Mid-ocean ridge and ocean-floor petrology*; *Ocean-island igneous rocks*; *Petrochemical calculations*; *Rift valley volcanism*; *Upper mantle—experimental petrology*.

BASIC IGNEOUS ROCKS

Because most igneous rocks are composed of silicate minerals, the earliest chemical classification used was one based on weight percentage of SiO₂ in the rock. This led to the subdivision of igneous rocks into four categories—acid(ic), intermediate, basic, and ultrabasic. Over the years, different authors have varied slightly in the limits of SiO₂ percentage for the four groups, but many petrologists designate igneous rocks with >66% SiO₂ as acidic, 52–66% SiO₂ as intermediate, 45–52% SiO₂ as basic, and <45% SiO₂ as ultrabasic (>65%, 65–55%, 55–45% and <45% are also used). Some authors use the term subsilicic and others use the term mafic

synonymously with basic although mafic is mainly used in relation to dark-colored Mg–Fe minerals or rocks rich in these minerals. The terms acidic and basic may also be used in a relative manner, e.g., to indicate that one rock is more basic than another even though the rock referred to may contain more than 52% SiO₂.

With chemical analyses available for only a very small proportion of rocks for which mineralogical information is available (particularly from thin-section study), the classification based on strict SiO₂ percentage tends to be replaced in practice by a classification based on mineral content. The term basic has accordingly acquired a special meaning from prolonged usage in petrology, to cover the gabbros, dolerites, microgabbros, and basalts (Hatch et al., 1972). The term *basite* is a general term for basic igneous rocks and metamorphosed basic igneous rocks.

Derivation of magmas that consolidate to form basic igneous rocks is from the mantle (see **Basalt**).

Coarse-grained basic igneous rocks (gabbros)

These plutonic rocks are found in almost any large basic intrusion, and especially in the major layered bodies like the Bushveld Complex or the Tertiary centers of W Scotland. The three principal minerals they contain are plagioclase (more calcic than An₅₀), pyroxene, and olivine. Depending on the proportions of each present, these rocks can be considered as variations between three main types—*gabbro* (plagioclase and clinopyroxene), *norite* (plagioclase and orthopyroxene), and *troctolite* (plagioclase and olivine) (Tables 1 and 2). Rocks of the gabbro–norite series are silica-saturated and may contain accessory quartz. Those of the gabbro–norite–troctolite series are undersatur-

ated; this is shown by gradual substitution of pyroxene by olivine.

Medium-grained basic igneous rocks (dolerites and microgabbros)

This group essentially covers the rock-type traditionally known as *dolerite* or *diabase*. Because of inconsistencies in the use of the latter term, *microgabbro* is used by some authors (Hatch et al., 1972). These rocks are hypabyssal, mainly occurring in dykes and sills. Mineralogically the rocks are similar to the gabbroic rocks. Plagioclase (An_{>50}) is ubiquitous and depending on the relative proportions of clino- and orthopyroxene, microgabbro, and micronorite are recognized. According to the degree of silica saturation, the following rock-types may also be recognized: quartz dolerite (quartz microgabbro), dolerite (microgabbro), and olivine dolerite (olivine microgabbro), quartz micronorite, micronorite, and olivine micronorite. In quartz dolerite, the over saturation is reflected in the presence of a quartz–feldspar or glassy mesostasis.

Although ophitic and subophitic textures are common in these rocks, particularly in dolerites, it is no longer considered to be a necessary feature for a rock to be classed in this group. Both ophitic and subophitic textures may also occur in gabbros and basalts.

Fine-grained basic igneous rocks (basalts)

The rocks of this group are predominantly found as extrusive lavas. Mineralogically they consist of plagioclase (An_{>50}) and pyroxene, with olivine also present in most. Opaque ore minerals are usually present in varying amounts. In contrast to gabbros and dolerites (microgabbros), porphyritic texture is common and

TABLE 1. Mineral composition of plutonic basic igneous rocks

Olivine ^a	Clinopyroxene dominant	Orthopyroxene dominant	Pyroxene ^a		
	Gabbro	Hypersthene gabbro	Augite norite	Norite	
10%	Olivine gabbro			Olivine norite	90%
30%	Troctolitic gabbro			Troctolitic norite	70%
50%	Gabbroic troctolite			Noritic troctolite	50%
70%	Augite troctolite			Bronzite troctolite	30%
90%	Troctolite				10%
100%					0%

^a The percentage figures for olivine and pyroxene refer to percentages of total mafic minerals present.

TABLE 2. Modal analyses of basic igneous rocks^a

	A	B	C	D	E	F	G	H	I	J
Minerals	Gabbro	Hyperssthene gabbro ^b	Norite	Olivine gabbro	Diallage troctolite	Quartz dolerite microgabbro	Dolerite microgabbro	Olivine dolerite microgabbro	Tholeiite ^b	Alkali olivine basalt
Olivine	—	—	—	6.3	23.0	—	2.1	22.6	—	14.0
Orthopyroxene	2.0	16.2	28.7	7.9	—	—	}	}	}	—
Clinopyroxene	33.0	26.8	1.0	30.4	6.0	21.3				24.9
Plagioclase	60.0	57.0	63.2	52.9	68.0	59.5	41.8	27.7	26.1	51.3
Opaque minerals	5.0	—	0.1	1.0	3.0	1.9	4.0	1.8	13.0	5.1
Quartz	—	—	—	—	—	14.6	0.2	—	26.4	—
Micropegmatite	—	—	0.1	—	—	2.7	1.0	1.5	7.1	—
Mesostasis	—	—	6.9	1.5	—	—	—	—	—	—
Others	—	—	—	—	—	—	—	—	—	4.7

^a A–I from Wilkinson (1967); J from Clark (1956).

^b Wt. %.

any or all three principal minerals (plagioclase, pyroxene, and olivine) may be found as phenocrysts and/or in the groundmass (Wilkinson, 1967; Hatch et al., 1972).

More than in the gabbros and dolerites (microgabbros), classification of this group has been based on chemical criteria rather than mineral assemblages. This has resulted in the distinction of two main types: (1) over-saturated *tholeiites* and (2) under-saturated *olivine basalts*. As originally defined, the *tholeiites*, like the quartz dolerites (quartz microgabbros), contain a silica-rich mesostasis as well as plagioclase, pyroxene and a little olivine. The use of the term *tholeiite* has now been extended to cover basalts varying from siliceous to olivine-bearing types, but all have significant quantities of Ca-poor pyroxenes. The alkali olivine basalts show their undersaturation by the presence of modal olivine and normative nepheline.

Syenogabbros and trachybasalts

Plutonic rocks that combine the characters of syenite and gabbro (*syenogabbros*) and those lavas having features of both trachytes and basalts (*trachybasalts-trachydolerites*) also occur. These rocks are basic in the sense of having plagioclase ($An_{>50}$) and pyroxene but are also alkaline in containing alkali feldspar or one of the feldspathoid minerals. Kentallenite, essexite, teschenite, crinanite, theralite, kyllite, and lugarite are syenogabbros. While it is more difficult to allocate named rock-types into the trachybasalt group, tephrite and basanite are two undersaturated varieties (Hatch et al., 1972).

F. WHYTE

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Cross-references: *Basalt; Dolerite; Gabbro and norite; Igneous rocks—classification and nomenclature; Plutonic rocks—classification and*

nomenclature; Pyrolite; Upper mantle—experimental petrology.

BATHOLITH

Batholith was first used by Suess (1901) to describe a "stock"- or "shield"-shaped igneous mass either intruded into country rock, or formed by the fusion of older basement. The usage today generally follows the definition of Daly (1933), viz., "typically a large, cross-cutting, subjacent intrusive mass, that is, one with no visible or clearly inferable floor of older, solid rock . . ." (Fig. 1). As such there is a gradation in scale from *stocks* with an area of $<10 \text{ km}^2$ to *bosses* ($<100 \text{ km}^2$) to regional masses greater than $1,000 \text{ km}^2$ in area. The term is sometimes spelt "bathylith" and a rarely used synonym is abyssolith. The term *pluton* is commonly used for a deep-seated (plutonic) intrusion, but generally for bodies considerably smaller than those referred to as batholiths.

Large batholiths rarely consist of one rock type, and few that have been mapped in detail consist of a single intrusion. Diversity can result either from magmatic differentiation of a single intrusion or repeated multiple intrusions along a particular belt.

The rock-types that form batholiths are as diverse as plutonic rock varieties, but most large batholiths in orogenic belts are composed of members of the calc-alkaline suite, i.e., granite, granodiorite, monzonite, tonalite, diorite, and gabbro. Table 1 compares the percentage rock composition of the Sierra Nevada Batholith (Fig. 2) with various parts of the Coastal Batholith of Peru (Fig. 3), which are typical of

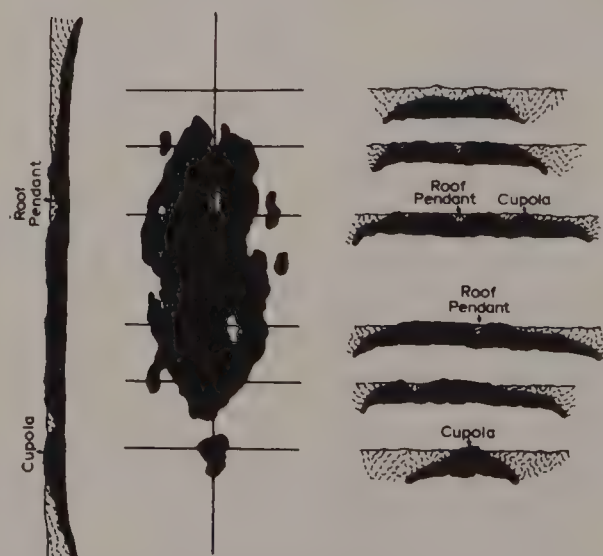


FIGURE 1. Plan and serial cross-sections illustrating the geometry of a batholith.

TABLE 1. Comparative percentages of rock-types in selected batholiths

Sierra Nevada ^a	%	Peru, southern ^a	%	Peru, central ^a	%
Tonalite	50	Tonalite	55	Tonalite–granodiorite	58
Granodiorite	34	Granodiorite–monzogranite	34	Monzogranite	25
Gabbro	14	Gabbro	7	Gabbro	16
Granite	2	Syenogranite	4	Syenogranite	1
Peru, northern ^a	%	Laitila, Finland ^b	%	Ahvenisto, Finland ^b	%
Tonalite	46	Granite–alkali granite	88	Granite–alkali granite	74
Granodiorite	20	Granodiorite–quartz diorite	8	Granodiorite–quartz diorite	13
Monzogranite	20	Anorthosite–gabbro	2	Anorthosite–gabbro	10
Gabbro	11.5	Syenite–mangerite	2	Syenite–mangerite	3
Syenogranite	2.5				

^a The Sierra Nevada and the Peruvian batholiths are syn- to late-orogenic plutons. (Data from Pitcher, 1978.)
^b The Laitila and Ahvenisto complexes are postorogenic rapikivi plutons in southern Finland. (Data from Savolahti, 1956, Vormaa, 1976.)

batholiths in an orogenic setting. In contrast, batholiths in other settings may be dominated by members of a granite–syenite trend or members of the gabbro–anorthosite–charnockite suite.

The setting of a batholith influences its overall

shape and appearance, and setting refers to the depth of emplacement as well as its tectonic position (Figs. 2, 4). Two concepts have evolved



FIGURE 2. Main batholiths of the Cordillera of western N America.

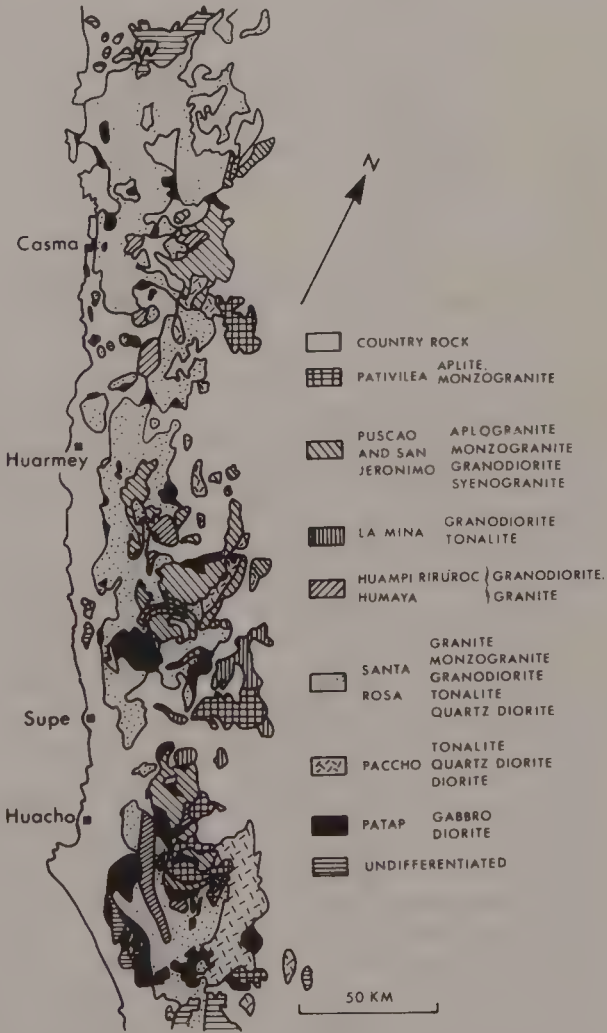


FIGURE 3. Diversity of rock types within a batholith complex, illustrated by part of the Coastal Batholith of Peru. (After Pitcher, 1978.)

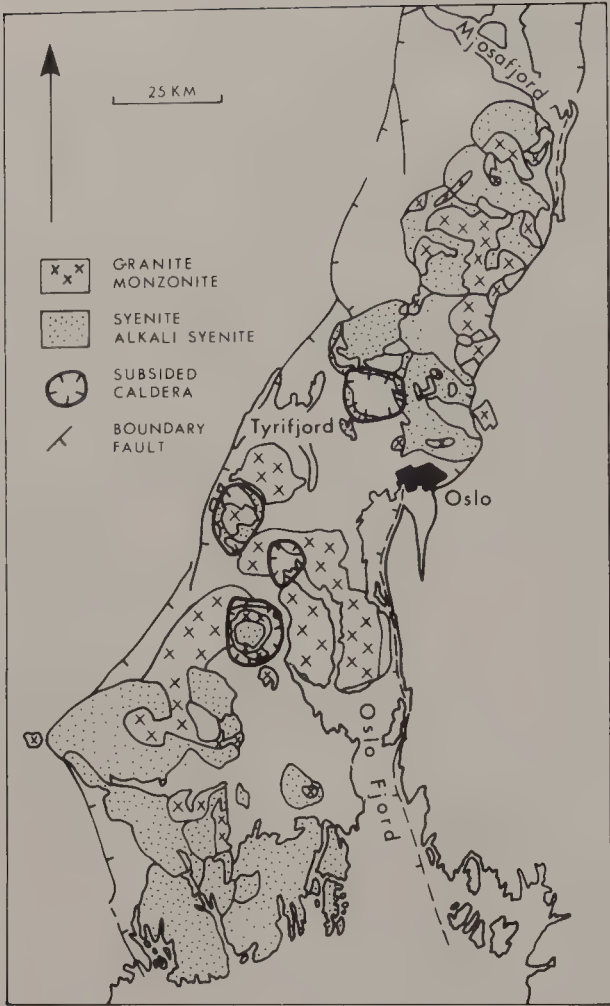


FIGURE 4. Batholiths, ring complexes and subsided cauldrons confined within the Oslo Graben, Norway.

to classify batholiths in these terms. Depth of emplacement is reflected in the terminology, *epizonal* (shallow emplacement), *catazonal* (deep emplacement) and *mesozonal* (in between), while the relationship to orogenesis and deformation is reflected in the terms *prekinematic* (pre-tectonic), *synkinematic* (syn-tectonic), and *postkinematic* (post-tectonic). *Anorogenic* is used for those batholiths not located in orogenic belts or not related to orogenic activity; to some extent the terms postkinematic and anorogenic overlap.

Although defined as discordant bodies, batholiths show a great variety of relationships to their host rocks, a variety that reflects depth of emplacement, mechanism of emplacement, and depth of denudation (Fig. 1). Some batholiths have sharp contacts with their host, others have gradational margins interpreted as vein injections, melting of the envelope, and metasomatic zones. In the case of the first two, some batholiths show a gradational relationship with migmatites and injection complexes; in the

latter case, the more fundamental issue of the origin of granites is raised and such processes as anatexis and granitization are invoked.

The actual emplacement mechanisms currently debated center on the properties of granitic magma. They fall into two categories—(1) forceful intrusion in which the magma shoulders aside the country rock, and (2) various permutations on stoping, in which masses of country rock founder into the rising magma. Forceful intrusion as a diapir has predictable consequences, in that the country rock should display the products of the deformation that has accommodated the rising diapir. Marginal schistosity in both batholith and aureole, and strain envelopes consistent with tangential stretching should be produced, while less extreme manifestations include rim-concentric folds and radial fracture patterns. Strain studies, particularly, have demonstrated the viability of the forceful emplacement model in the case of late tectonic granites (e.g., the Ardara granite in Donegal, Eire (Fig. 5) and the Sierra Modena granites, Spain—see **Diapirism**). While this approach has produced generally acceptable conclusions in the case of post- and late tectonic granites, in the case of preorogenic granites, where the batholiths are themselves deformed, often consisting of augen gneiss, diapiric models have been more controversial. The controversy here centers on the viability of models invoking the plastic mobilization or melting of granitic gneissose basement such that gravitational instability causes diapirs to rise into the overlying and generally denser cover. The mantled gneiss domes of eastern Finland have

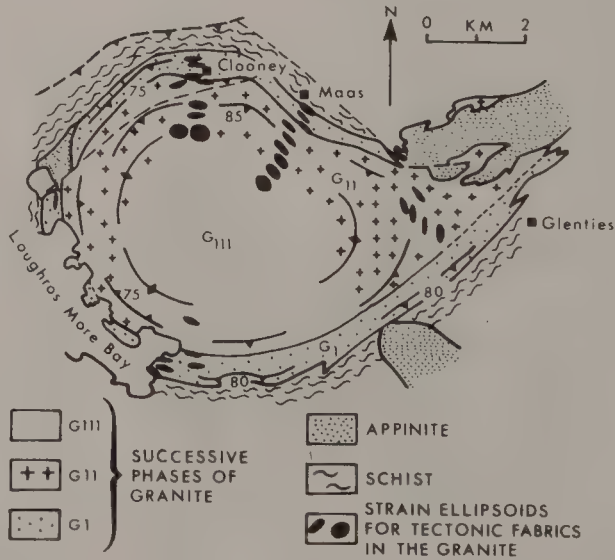


FIGURE 5. Strain ellipsoid shapes within and around the Ardara pluton, Eire, suggesting a diapiric emplacement mechanism. (After Holder, 1979.)

been regarded as classic examples of such a process. However, the interpretation of the regional pattern in such terranes has produced much controversy (Snowden and Bickle, 1976) and at least some "mantled gneiss domes" have been shown to be fold-interference structures.

Stopping models have also proved controversial, although debate has centered on the scale of the process rather than its fundamental viability. Piecemeal stopping is indicated by the existence of country rock xenoliths, often forming intrusion-breccias, in the margins of many batholiths. Such xenoliths may pick out a "ghost stratigraphy," e.g., in the Donegal Granite, Eire (Pitcher and Berger, 1972). "Ghost stratigraphy," especially where assimilation of xenoliths can be invoked, has been used as evidence of massive granitization. However, most workers now agree that although such assimilation does occur, it results in contamination of an invading magma, rather than reflecting wholesale anatexis in situ.

Stopping on an altogether larger scale is now invoked to explain the emplacement of many mesozonal and epizonal granitoids. This emplacement mechanism results in the development of ring dykes: large rafts of country rock founder like pistons into a rising,

passively-emplaced granitoid batholith. Such a mechanism relates directly to Glen Coe-type cauldron subsidence, and such batholiths, often occurring as contiguous swarms (Figs. 3, 4, 6), have a generally circular plan, including massive ring dykes, circular screens of country rock, and foundered masses of cogenetic volcanic rocks, often the products of explosive volcanic activity. Examples include the ring complexes of Nigeria, the granites and syenites of the Oslo Graben, the granites of the British Tertiary Volcanic Province, and the various granitoids and volcanic basins of Peru.

The stopping model should not be seen as exclusively passive and distinct from forceful intrusion. All batholith emplacement involves a space problem and the two models are opposite ends of a spectrum devised to cope with it. Furthermore, examples of forceful stopping have been documented (Ehlers and Bergman, 1984). In other areas the various depth zones are superimposed in such a way that epizonal batholiths in one province show a relationship to ring dykes and subsided cauldrons, while cogenetic bodies exposed at deep levels show marginal migmatites and anatectic features. Some migmatite complexes have been described as the roots of a batholith (cf., Kenah and Hollister, 1983).

ADRIAN F. PARK

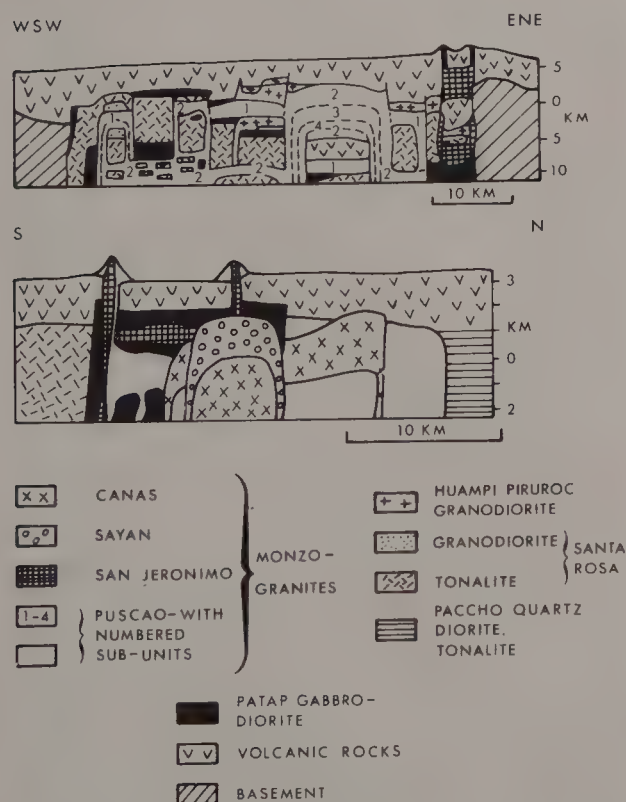


FIGURE 6. Schematic relationship between various units of the Coastal Batholith of Peru and contemporaneous volcanic rocks, suggesting a stopping and subsidence emplacement mechanism. (After Pitcher, 1978.)

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Cross-references: *Anatexis*; *Aureole*; *British Tertiary Volcanic Province*; *Calderas and volcano-tectonic depressions*; *Charnockite suite*; *Diapirism*; *Emplacement—modes*; *Granite*; *Igneous intrusions—structural behavior*; *Injection complex*; *Intrusion*; *Migmatite*; *Ring dyke*; *Xenolith*.

BRECCIA PIPE

Breccia pipes are cylindrical or irregular bodies of brecciated rock that may show cross-cutting or gradational relations to the country rock. They are sometimes associated with igneous bosses or volcanic pipes, or show extensive mineralization. Others have no evidence of associated igneous or hydrothermal activity. Breccia pipes are almost certainly polygenetic and include collapse-breccias formed by subsidence, explosion-breccias formed by explosion of magmatic gases, and intrusive breccias formed by the mobilization of explosion-breccias (Bowes and Wright, 1961; Wright and Bowes, 1968). The first two types have many features in common. The blocks are normally disoriented and may be of mixed origin. There is very often a gradational zone of shattered but unmoved rock around the breccia pipes that may be of considerable width, depending on the strength of the country rock. The matrix of the breccia is of comminuted rock similar to that forming the phenoclasts, except in those cases where igneous or hydrothermal material has later pervaded the breccias (Bryner, 1961).

The collapse-breccias are very frequently associated with igneous bodies, e.g., in the Cananea pipes, Mexico (Perry, 1961), and in these and other cases the downward displacement of the country rock can be proved and is indeed essential to the proof of collapse as the mechanism of brecciation. Radial and dome-shaped fracture patterns are typical of this type of breccia. These breccias are thought to be due to gravity collapse on withdrawal of magma during oscillatory intrusion of subvolcanic plutons. Mineralization, which is common, is always later than the brecciation.

Other collapse-breccias have no associated igneous bodies but may still be heavily mineralized. The subsidence in these cases may be related to solution by hydrothermal fluids whether there are deposits in the interstices of the breccia or not. Vugs and open stockworks are typical features of collapse-breccias that are absent from explosion-breccia pipes.

Explosion-breccias are normally associated with igneous bodies. A gradational from igneous rock to breccia, by intrusion of the igneous rock into the interstices of the brecciated blocks, is usual. The gradation from breccia to country rock is also apparent in true explosion-breccias but they are frequently somewhat mobilized into intrusive breccia masses and in this environment the margins are quite sharp. Upward transport of boulders and the mixing of angular and rounded fragments, some extremely rounded, is a characteristic of intrusive breccia pipes and they grade towards the surface into volcanic diatremes. The boulders may be transported upwards many thousands of meters but in other cases the upward termination of the pipe may be present. This type of breccia is generally associated with particular types of igneous rock, the following petrogenetic provinces being notable for their associated breccias: (1) alkaline and carbonatitic volcanoes particularly of eastern and southern Africa, (2) kimberlites, (3) quartz porphyries and monzonites together with latite sills and dykes of the western United States, (4) appinites, kentallenites, diorites, and syenites of the Caledonian orogenic belt of western Scotland and Northern Ireland, (5) acidic rocks associated with the basaltic British Tertiary Volcanic Province of western Scotland and northeastern Ireland, and (6) alkaline lamprophyres in many parts of the world.

In these cases there is a strong correlation between the brecciation and the presence of extreme differentiates that are rich in both volatiles and alkalis. It is this close correlation with highly volatile magmas that has led to the use of the term explosion-breccia, although whether it is possible for subsurface volcanic explosions to take place is still debatable. The explosions are generally ascribed to supercooling and rapid crystallization causing very great increases in vapor pressure (Morey, 1922).

Another mechanism in which the very high gas pressures are essential is described by Hardie (1963); in this the gas forces apart fractures in the rocks and it is the sudden release of gas upwards when an open system is reached that causes the fractures deeper down to be relaxed so suddenly as to cause shock waves that result in brecciation of the rock. Pitcher and Berger (1972) consider the brecciation of similar bodies in Donegal to be tectonic and only the transport to be due to the volcanic activity, but there is an almost universal occurrence of breccia pipes with highly volatile magma.

Mobilization of breccias associated with gas action leads to their intrusion higher in the crust by the process of fluidization, possibly up

preexisting zones of weakness, to form pipes of intrusive breccia. In some cases movement of quite large blocks can be shown to be downwards as well as upwards (Coe, 1966). The location of breccia pipes can often be shown to be controlled by preexisting structures. Some are situated where gas has been trapped in anticlinal vergences in areas of multiple folding; in others the intersections of joints or fault lines are utilized as a means of escape of gases (Bowes and Wright, 1967). In all cases the formation of the breccias is in a confined space underground and breccia pipes are thus dissimilar to volcanic necks.

A. E. WRIGHT

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Cross-references: Appinite; Carbonatite; Diatreme; Explosive volcanic eruptions—classification; Fluidization; Kimberlite; Volatiles; Volcanic neck.

BRITISH TERTIARY VOLCANIC PROVINCE

The British Tertiary Volcanic Province (BTVP) is part of a larger N Atlantic Tertiary Province that includes NW Britain, E Greenland, the Faroes and Iceland (see **Flood basalt**). The igneous activity in this major province is related to the plate separation and subsequent sea-floor spreading between Greenland and NW Europe. NW Britain, E Greenland and the

Faroes are situated on continental crust, but Iceland is of truly oceanic character and represents the site of current igneous activity on the spreading ridge axis that bisects the island. The structural setting of the igneous activity in the BTVP differs from that of E Greenland and the Faroes in that it lies within the continental crust while E Greenland and the Faroes are situated at the margin adjacent to ocean crust. The voluminous effusion of tholeiitic basalts in E Greenland and the Faroes immediately preceded plate separation. In the BTVP, alkaline basaltic and transitional magma types and their derivatives occur in association with the tholeiitic magmas formed at the climax of igneous activity.

Radiometric dating indicates that the igneous activity in the BTVP (55–61 Ma) is slightly older than that of E Greenland and the Faroes, with the peak of activity being around 57 Ma ago.

The principal elements of the BTVP are plateau lavas, central igneous complexes, and regional dyke swarms (Figs. 1, 2). The relationships between these elements are still imperfectly understood. The major lava fields usually predate the central intrusive complexes. Dyke emplacement occurred throughout the evolution of the central intrusive complexes and some dykes are contemporaneous with the lavas, while others cut all of the currently exposed lavas. The regional dyke swarms focus



FIGURE 1. Positions of central complexes, lava fields and dyke swarms; additional submarine and on land (Rockall) central complexes are further west. (After Emeleus, 1983.)

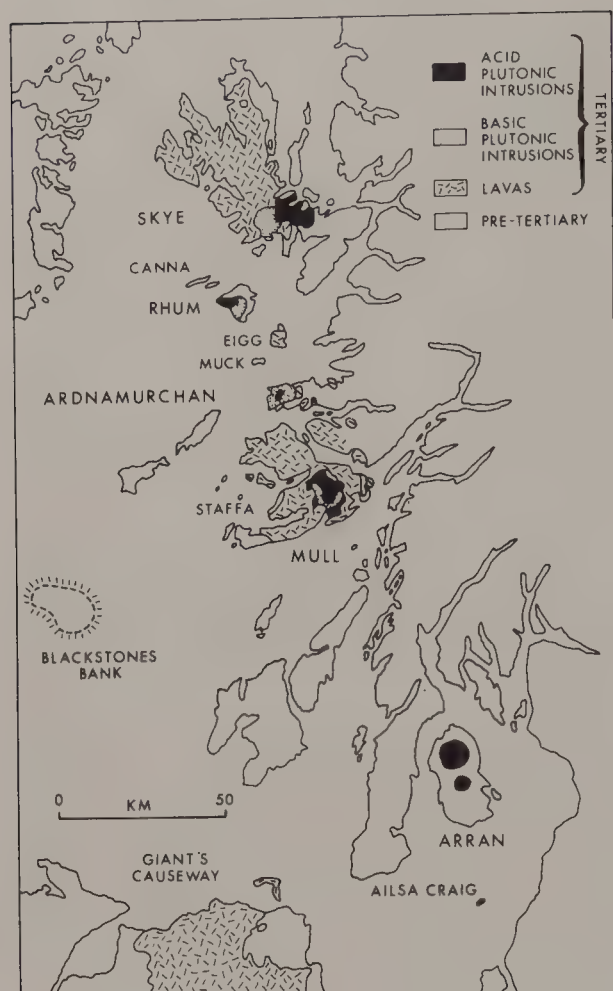


FIGURE 2. Distribution of lavas and plutonic intrusions. (After Sutherland, 1982.)

on some of the central complexes and it has been suggested that such complexes developed where the en echelon linear ridges of magma that fed the dyke swarms intersected major lines of crustal weakness.

The BTVP has been the scene of many pioneering petrological studies, and its features have formed the basis for many important concepts in the geology of igneous rocks, particularly relating to subvolcanic, shallow-level igneous intrusions. In petrology, many of the early concepts of magmatic fractionation (acidic and basic rocks coming from one parent), hybridization, and the effects of liquid immiscibility, were formulated through these studies. The recognition of ring dykes and cone sheets, and ideas regarding their mode of emplacement, reflect the excellent development and exposure of the subvolcanic complexes on the Isle of Mull and Ardnamurchan (see **Cone sheets** and **Ring dyke** for maps of Ardnamurchan and Mull, respectively). More recently, the understanding of processes operative in shallow, basic magma chambers, namely, the role of crystal fractionation and double-diffusive con-

vection, has been enhanced particularly by studies of the layered intrusions of Rhum and Skaergaard (in the coeval E Greenland Tertiary province). Recent geochemical and isotopic studies in the BTVP have led to an appreciation of the complexity of synergistic influences, namely, mantle source heterogeneity, mantle melting, crustal anatexis and assimilation, and magma hybridization, in the evolution of a single igneous province (see Thompson, 1982, for references). The individual complexes and the overall nature of the BTVP have been described in detail in reviews by Sutherland (1982) and Emeleus (1983), where full references to the wealth of literature can be found. The petrogenetic arguments of recent times are reviewed by Thompson (1982).

Plateau lavas

Remnants of an extensive, thick accumulation of predominantly basaltic lavas crop out in NE Ireland, the Scottish Hebrides from Mull to Skye, on the adjacent Scottish mainland, and on the surrounding sea floor (Fig. 2). The current area of outcrop is approximately 7,200 km² and the thickness of the lava pile varies between 1,800 m on Mull, 800 m in NE Ireland and 600 m on Skye and the adjacent island of Raasay. The lava flows are generally amygdaloidal and the original vesicles are now infilled by zeolites, various forms of silica, and calcite. Studies of the zeolite zonation suggest that a further 400 m of lavas have been removed by erosion.

The lavas were erupted subaerially on to an uneven terrain, although minor developments of pillow lava and hyaloclastite indicate local subaqueous eruptions. Lateritic soils, or "boles," are frequently developed on top of some lava flows, and terrestrial plant remains occur in some interbasaltic sediments: fossils trees are present within one lava flow on Mull. The lavas appear to have erupted mainly from fissures, though at least some of the flows on Mull and Skye relate to the central volcanoes.

Many compositional and textural varieties of lava are present. The lava succession on Mull was subdivided into a thick lower group of olivine basalts, termed the "Plateau Basalt-type," and an upper group of aphyric basalts termed the "Non-Porphyrific Central Basalt-type." These two magma types were subsequently recognized to be nepheline-normative alkali basalts and hypersthene- or quartz-normative tholeiitic basalts, respectively. In the light of more recent geochemical studies, a greater diversity of lava types has been recognized and the importance of silica-saturation as a discriminant between basalt

types has diminished. The following magma-types have been recognized on Skye and Mull and petrogenetic schemes to account for them have been proposed.

1. *Skye Main Lava Series (SMLS)*—mildly alkalic olivine tholeiites parental to a hawaiite–mugearite–benmoreite suite by deep fractionation of clinopyroxene just beneath the crust; also parental to trachytes when fractionation has occurred at higher oxygen fugacities at higher crustal levels.
2. *Preshal Mhor (PM) basalts*—also olivine tholeiites but less alkalic and more calcic than the SMLS type; derived by a second melting episode of the mantle depleted in large-ion lithophile (LIL) elements by the formation of the SMLS melts, leaving a harzburgitic residuum.

The more fractionated flows become commoner higher in the succession. Flows of the MORB-like PM basalts occur near the top of the succession and dykes of a similar composition occur in the axial regions of the regional dyke swarms. Another magma-type, the *Fairy Bridge (FB) type*, occurs within the regional dyke swarm and represents a rare intermediate between SMLS and PM types.

The plateau basalts of Mull are similar but not identical to the SMLS basalts. Several distinct types have been recognized.

3. *Mull Plateau Group I*—mildly alkaline basalts, hawaiites, and mugearites with a distribution pattern of Y and Sr suggesting derivation from a garnet-bearing source.
4. *Mull Plateau Group II*—less alkalic basalts with derivative trachytes and Sr:Ba ratios suggesting an origin from plagioclase-bearing lherzolite.
5. *Mull Plateau Group III*—a possible mixture of MPG I and MPG II, or derived from a source with intermediate properties.
6. *Staffa magma-type (?)*—high-Ca–low-alkali tholeiites have been recognized in the axial regions of the Mull regional dyke swarm.

Other local variations in basalt chemistry occur within the different parts of the BTVP, reflecting mantle heterogeneities or different melting episodes, or indicating different fractionation histories.

Acidic lavas are rare in the province. Icelandites are preserved on Rhum and effusive rhyolite forms domes and flows at Tardree in NE Ireland, and a flow on Eigg.

Central intrusive complexes

The central intrusive complexes of the BTVP represent plutonic rocks formed at localized

centers of igneous activity beneath central volcanoes. They are characterized by major gabbroic and granitic intrusions with explosion vents, multiple cone sheets, ring dykes, plugs, sills, inclined sheets, and local dykes. In each complex there is a considerable diversity in rock type and mode of emplacement, although a marked bimodality of acidic and basic rock types, with a relative absence of intermediate compositions, is a persistent feature.

The central complexes are typically 10–15 km in diameter and geophysical evidence suggests that each is underlain by subvertical, cylindrical masses of dense basic and ultrabasic rock that extend down to at least 15 km, and probably into the mantle. Each complex is usually formed by a number of arcuate intrusions that define one or more focal points for intrusive activity.

The proportions of acidic and basic rocks at the current level of exposure varies throughout the BTVP. The Arran complex is predominantly acidic although underlain by denser basic rocks, while the Ardnamurchan complex is predominantly basic.

Relationships of acidic and basic magmas

Within the central complexes of the BTVP there are several striking examples of the simultaneous existence of acidic and basic magmas. These take the form of composite intrusions, hybrid rocks, gradational intrusions, and net-veined complexes.

Composite sills and dykes, such as the Drumadoon and Bennan Head sills in Arran, typically show basaltic margins and quartz-feldspar porphyry interiors. The acidic rock is not normally chilled against the basic member and complex reaction relationships can often be observed. In some sills a hybrid zone has formed between the marginal basalt and central felsite.

Hybrids that show greater overall homogeneity form sills and dykes throughout the province. Such hybrid rocks are frequently called *craignurite* after the hybrid sill at Craignure on Mull. In some of the ring dykes the mixing of magmas with different compositions and temperatures has produced mixed magma rocks and net-veined complexes. The mafic components in these cases are rarely basalt and more typically basaltic andesite, andesite or dacite. The Marsco complex on Skye includes the intermediate hybrids marscoite and glamaigite, which represent a mixture of ferroandesite (ferrodiorite) and rhyolite magma. In other ring dykes, such as on St. Kilda, pillow-like bodies of mafic and intermediate rock occur in a granophyric host. The basic pillows are often chilled against the acidic host

and may show crenulate margins. Elsewhere, the basic rock can be intimately veined by the acidic member, as seen in Centre 2, Ardnamurchan. The Loch Ba ring dyke on Mull provides an example of more complete mixing of magmas. It contains wispy inclusions of basaltic andesite and andesite in a host of micro-crystalline felsite.

The mixed magmas and net-veined complexes are considered to have formed by the violent emplacement of thermally and compositionally zoned magma systems owing to collapse of a subvolcanic magma chamber, resulting in incomplete mixing of magmas with contrasting properties from different levels of the chamber.

Origin of the acidic magmas

The origin of the acidic rocks has been the topic of great debate. Early ideas revolved around origins by either crustal melting or differentiation of a basaltic parent magma. Recent isotopic and trace-element studies have shown that crustal melting, contamination, remelting of existing Tertiary granitic rocks, and fractionation may all have occurred to varying degrees to produce the different granitic rocks. The earlier melts in both Skye and Mull tend to be richer in crustal components than the later granites, which have higher ratios of mantle-derived magma to crustal melt.

Sr, Nd, and Pb isotope data indicate variable but often considerable fractions of amphibolite facies gneisses of the Lewisian complex in the Skye and Mull granites, in contrast to the mostly granulite facies contaminants in the preceding lavas. Arkoses of the late Proterozoic Torridonian assemblage have been confirmed by recent isotope studies to be the source of a granophyre on Skye.

Emplacement and petrogenesis of the basic intrusions

The major basic and ultrabasic intrusions in the BTVP are typically arcuate in form, though not strictly ring dykes. Many intrusions are relatively homogeneous in composition with little internal structure, but others display marked rhythmic layering. The layered intrusions of Rhum, Skye, Mull, Ardnamurchan, and Carlingford show lithologically distinct units that vary in scale from centimeters to tens of meters. Lithological differences are marked by different amounts of magnesian olivine, calcic plagioclase, diopsidic augite, and chrome spinel; in the case of the layered series on Rhum there is an alternation of layers of peridotite and allivalite (Ca-rich anorthosite). Cryptic variations in the composition of these constituent minerals have been detected in some cases and

in conjunction with trace-element studies; these can be used to interpret mixing processes within the magma chamber. The lithological variation has been attributed conventionally to processes of crystal settling from a convecting magma system. The formation of ultrabasic breccias, slumping of layers, and the formation of erosion channels testify to the former existence of an active environment for crystal accumulation. Unusual rock textures are associated with "diagenetic" changes in the magmatic sediment and the spectacular harrisites of Rhum result from the rapid growth of olivine from a hydrous peridotite magma.

The major basic intrusions of the BTVP seem to have formed from a Skaergaard-type tholeiitic basalt enriched in the light rare-earth elements. There has been some debate whether this parental magma was ultrabasic or basic, and both feldspathic peridotite and basalt are still cited as possible parental compositions to some of the major layered intrusions. The intrusion of a suspension of olivine crystals in an ultrabasic magma has been proposed to produce the layered successions of Rhum and Skye.

Regional dyke swarms

Extensive NW-SE-trending linear dyke swarms are closely associated with the central complexes, but also occur independently. The dykes are principally basaltic and it is only in the immediate vicinity of the central complexes that intermediate, acidic and composite dykes occur. The chemical composition of the dyke rocks varies spatially, with low-alkali tholeiites occurring in the axial regions where crustal dilation is highest. More alkaline types occur off-axis and in some of the smaller swarms where crustal dilation was less.

Along the axes of the individual swarms, the crustal dilation may be locally very high. In southern Skye, adjacent to the central complex, maximum crustal dilation is 25%, represented by a dyke density of nearly 300 dykes/km. In the case of the Mull dyke swarm, the maximum dilation was only 10%, with a maximum dyke density of 70 dykes/km (Fig. 3).

The dykes have been emplaced upwards from linear ridges of magma that formed in response to pressure release associated with the propagation of major tensional fractures in the crust. This may have been related to the shearing stress associated with the opening of the N Atlantic Ocean. It seems likely that the central igneous complexes formed where these magma ridges intersected major lines of crustal weakness, which focused the diapiric rise of the major bodies of magma.

G. P. DURANT

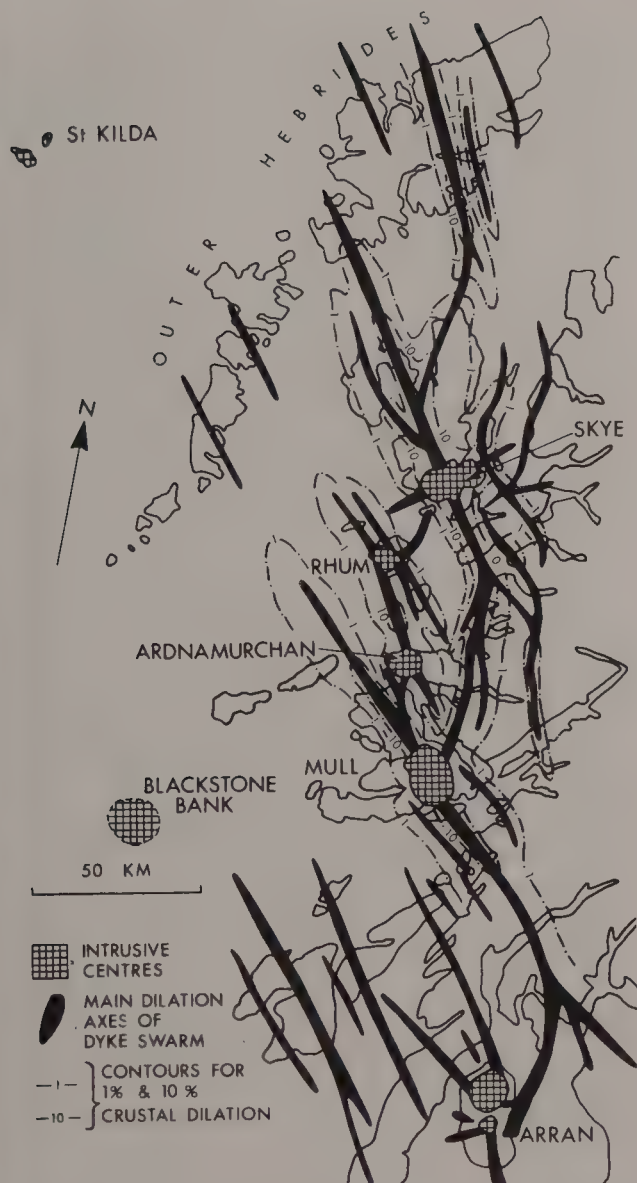


FIGURE 3. Crustal dilation in relation to dyke emplacement. (After Sutherland, 1982.)

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Cross-references: *Cone sheets*; *Cumulate textures*; *Flood basalt*; *Hybridization*; *Igneous rocks—tectonic setting*; *Intrusion—layered*; *Ring dyke*; *Skaergaard intrusion*.

* Contains an extensive reference list.

BROWNSTONE FACIES

The *brownstone facies* was proposed by Cann (1979) as a low-temperature mineral facies encompassing ocean-floor weathering and cool hydrothermal alteration on the ocean floor. The most widespread secondary minerals under oxidizing conditions include a K-rich dioctahedral Fe-illite resembling celadonite. This replaces olivine, occupies vesicles, replaces interstitial glass, and may eventually replace augite. Under reducing conditions, its place is taken by saponite, and pyrite is also characteristic. Plagioclase may be replaced by clay minerals or potassium feldspar. Glassy rinds of basaltic pillow lava alter to palagonite in association with phillipsite and other low-temperature zeolites and calcium carbonate.

D. S. COOMBS

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Cross-references: *Metamorphic facies*; *Ocean-floor metamorphism*; *Palagonite*; *Spilite*.

BUCHITE

Buchite is a very uncommon hornfels containing glass. The term was first used by Mohl (1873; see Spry and Solomon, 1964), apparently named after Leopold von Buch (Tomkeieff, 1940; see Spry and Solomon, 1964). The presence of a glass phase implies partial melting during metamorphism. Typical minerals include cordierite, mullite, tridymite, spinel, corundum, pyroxene, and feldspar (particularly sanidine), generally as very tiny crystals within an alkali–alumina–silica glass that also contains unmelted relics of original minerals (Fig. 1). Subspherical bodies of opal or chalcedony are common and appear to be amygdulites, indicating an excess gas pressure during melting.

Buchite occurs adjacent to volcanic necks (Apsley, Tasmania; Owens Valley, California; Sydney, N.S.W.), near basic intrusions (Heilbron, Orange Free State; Ardnamurchan, western Scotland), as inclusions in volcanic rocks or flows, or near burnt coal (Jharia, India), and the varieties described have been derived from sandstone, shale, breccia, and granitic rocks.

Columnar jointing is not uncommon. The physical conditions for the formation of buchite

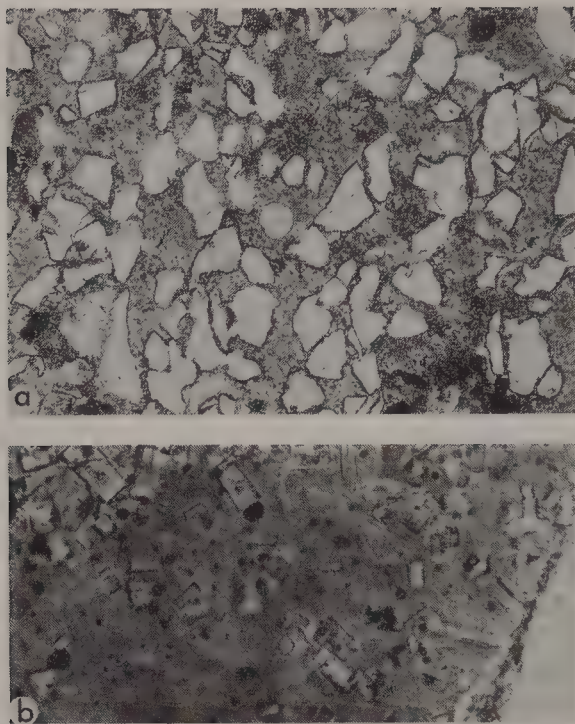


FIGURE 1. Photomicrographs of buchite. (a) Corroded grains of quartz set in light brown glass with dark aggregates of spinel and cordierite; PPL, $\times 3$. (b) Relict quartz fringed with tridymite and surrounded by glass containing euhedral cordierite; PPL, $\times 60$.

appear to be a temperature in the vicinity of 900–1,100°C and very low pressures (ca. 0.1–0.5 kb). Such extreme conditions are most likely to be achieved adjacent to volcanic necks through which magma has flowed for an extended period or where convective heat transfer has taken place by passage of magmatic water vapor through permeable sediments.

ALAN SPRY

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Cross-references: *Aureole*; *Contact (thermal) metamorphism*; *Hornfels*; *Pyrometamorphism*; *Pyrometasomatism*; *Volcanic neck*.

BURIAL METAMORPHISM

Burial metamorphism is here equated with *depth* or *load metamorphism*. The term was first used by Coombs (1961) to describe a

process of metamorphism that resulted in large-scale recrystallization but with no strong penetrative deformation. Its products occur in large areas in New Zealand, Japan and elsewhere, and it defines the lowest stage of metamorphism. Ocean-floor metamorphism has many similar characteristics. Not uncommonly there is a gradation with decreasing metamorphism from greenschist facies rocks to rocks containing prehnite and/or pumpellyite, then into rocks containing zeolites, and finally into unrecrystallized rocks. As rocks that have suffered burial metamorphism can be distinguished readily from ordinary unmetamorphosed equivalents, there is clearly a difference between diagenesis and burial metamorphism. Nonetheless, many of the reactions and products are the same, so that a precise fixing of the points at which diagenesis ceases and burial metamorphism starts is difficult. In some senses there has been too much emphasis, unconsciously at least, on the continuity of diagenesis and depth metamorphism. In fact, diagenesis is usually complete long before burial metamorphism sets in, and the latter occurs at various depths depending on composition as well as physical variables other than pressure ($\equiv$ depth). Nevertheless, some of the processes are the same: for instance, the recrystallization of clay minerals and the formation of carbonates. Characteristically, rocks that have undergone burial metamorphism show their original texture and a mineralogy that has only been partly modified; new minerals are often confined to interstices, veins, vesicles, or alteration zones.

Facies

The initiation of metamorphism under conditions of burial metamorphism is very variable and particularly sensitive to rock and fluid composition and to the geothermal gradient, so much so that there is considerable variation in the assemblages from one area to another. The start of metamorphism must therefore be represented by a wide field in P – T – X space.

The two metamorphic facies of burial metamorphism (Coombs, 1961) are, with decreasing grade following the greenschist facies, (1) prehnite–pumpellyite facies and (2) zeolite facies (see also **Pumpellyite–actinolite schist facies**). These facies are characteristic of extensive areas of eugeosynclinal sediments, usually graywackes, lavas, and pyroclastic materials, which are particularly sensitive to recrystallization at low temperatures. Following Miyashiro (1973) and Turner (1981), burial metamorphism can be considered in terms of

* This paper contains references to earlier work cited.

different P - T paths or geothermal gradients. In regimes with higher geothermal gradients large numbers of zeolites are produced, e.g., stilbite, mordenite, yugawaralite, wairakite, heulandite, analcime, and laumontite; where the geothermal gradient was lower, only the last three or perhaps just the last mineral may be produced.

A good example of low-pressure-high-geothermal-gradient burial metamorphism is seen in the rocks of the Tanzawa Mountains, Japan where the metamorphism varies from zeolite to amphibolite facies (Seki et al., 1969). The low-grade rocks are undeformed and made up of andesitic and basaltic material. Metamorphism was a long process, for it started in the early Miocene rocks (Tanzawa Group) and continued in younger rocks of the late Miocene and Pliocene (Ashigara Group). The zonal sequence up to greenschist facies rocks is shown in Fig. 1. Incomplete recrystallization is common at least to zone 3, while original textures are preserved beyond zone 3. Depths postulated by Seki et al. (1969) are 5–8 km, with a geothermal gradient about 40–60°C/km.

In many active geothermal fields a similar sequence to the above has been noted, although rocks of the prehnite–pumpellyite facies zone are absent. The geothermal gradients are higher than those of the low-pressure type (200–1,000°C/km). However, in some geothermal

fields zeolite formation is absent or markedly restricted, probably owing to solutions containing abundant CO_2 as well as other components, emphasizing the importance of fluid composition, especially the $\text{H}_2\text{O}/\text{CO}_2$ activity ratio, on the assemblage formed.

The type area for the zeolite facies is Taringatura, New Zealand (Coombs, 1961), which is characterized by a lower geothermal gradient than the above areas (ca. 22°C/km—Miyashiro, 1973). Stilbite, mordenite, wairakite, etc., are not found. The Triassic sediments showing burial metamorphism are some 10 km thick, forming part of the southern New Zealand eugeosyncline, and are made up of graywackes and acidic tuffs. Two assemblages are noted in the zeolite facies, characterized by heulandite–analcime and laumontite. With increasing depth, and with no deformation, the grade of metamorphism increases to prehnite–pumpellyite facies.

The examples above relate to volcanic and volcanoclastic sequences, but other rocks such as shales, marls, and graywackes also show the effects of burial metamorphism, although it is usually more subdued. For example, Frey (1970) described the metamorphism in shales and slates in the Alps. Minerals characterizing burial metamorphism in these rocks include illite, chlorite, pyrophyllite, paragonite–muscovite, paragonite, stilpnomelane, and rarely chloritoid. In such rocks, illite crystallinity, as shown by the first-order basal reflection (Kubler, 1967), has been used to indicate grade (cf., Oliver et al., 1984). Additionally, coal rank determinations on carbonaceous material by reflectivity techniques and fluid inclusion studies to determine the composition of metamorphic fluids as well as temperature and pressure of entrapment of fluid have been used to establish areal variations and intensity of burial metamorphism. Thus, in graywackes, pelites, and marls of the external parts of the Alps, Frey et al. (1980) have demonstrated that grade increases from tectonically higher to lower units and from the external to the internal parts of the same tectonic unit.

Figure 2 shows approximate correlations of various parameters with increasing grade. It serves to indicate the style of approach in studying this type of metamorphism, but should be viewed with caution considering the present lack of understanding of the variables controlling the assemblages.

Limits of burial metamorphism

Although experimental data exist for some the minerals involved (Zen and Thompson 1974; Turner, 1981), there are little or no data

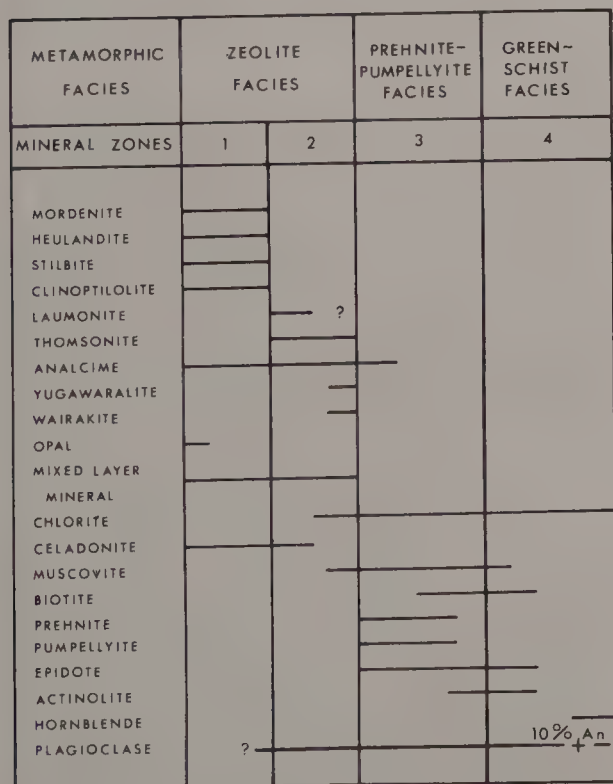


FIGURE 1. Distribution of minerals in basic and intermediate rocks in the Tanzawa Mountains, Japan. (After Seki et al., 1969.)

BURIAL METAMORPHISM

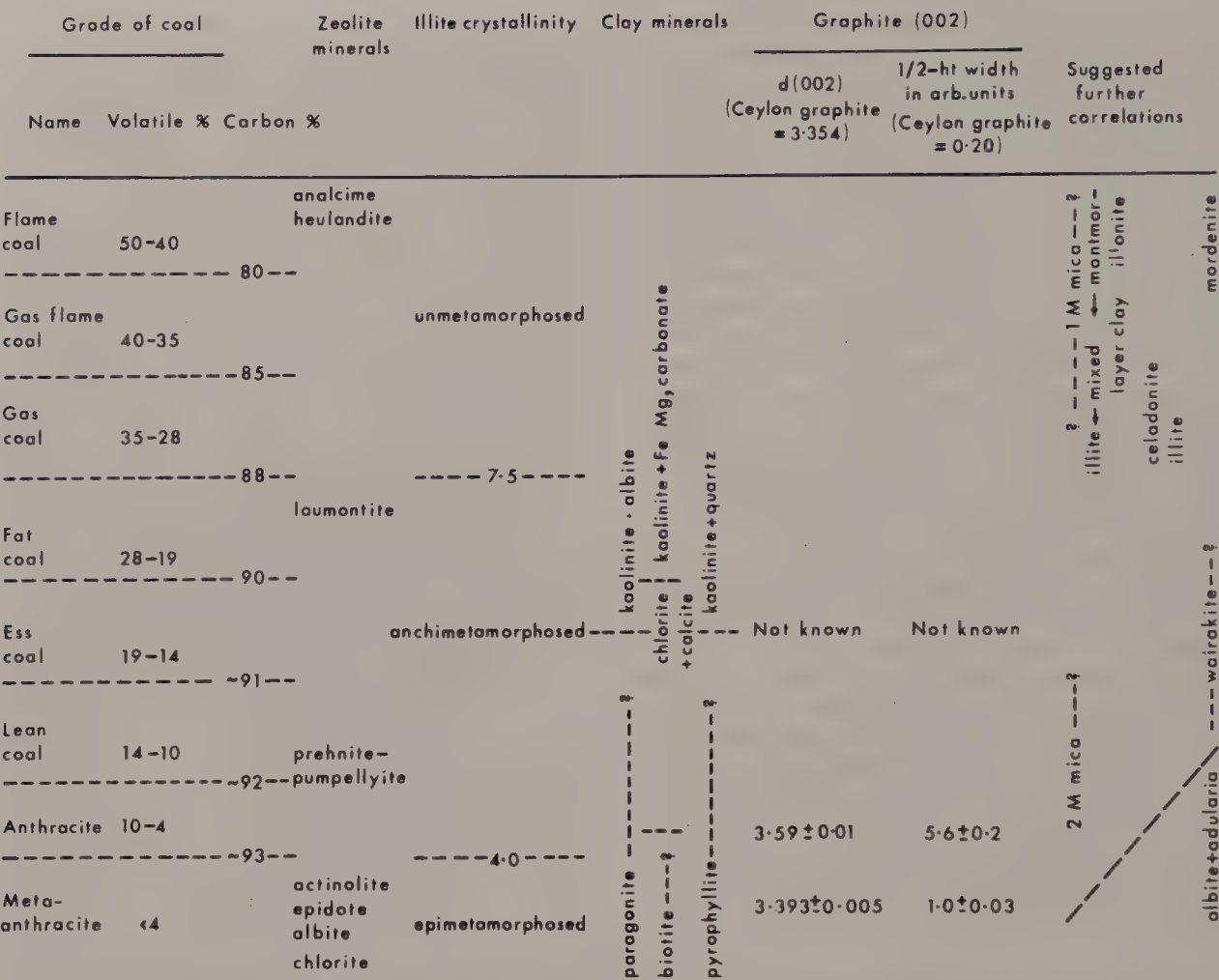


FIGURE 2. Approximate correlations of coal rank, zeolite minerals, illite crystallinity, clay mineralogy, and graphite crystallinity. (After Zen and Thompson, 1973.)

on the stability fields of some important assemblages, e.g., pumpellyite-prehnite; neither are there good data on H₂O-CO₂ equilibria for any of the major "zone markers." The problem of limits is further compounded by the clear lack of equilibrium and possibility of metastability in these rocks. Accepting this and using field and textural evidence, tentative conditions for burial metamorphism lie below 3 kb and 350°C, with wairakite assemblages occurring below 1 kb and between 200° and 280°C.

Implied in the above discussion is the simple concept of burial metamorphism as a function of depth, variations being accounted for by different geothermal gradients in different areas. However, in the "eugeosynclinal" environment, zeolitic assemblages may be confined to vesicular domains and tops of lava flows (Offler et al., 1980). Repeated patterns as seen in Peru and Chile in which grade increases rapidly in each "unit" are also characteristic of this marginal basin setting. Here the intensity of metamorphism and repetition appear to relate

to a high heat flow within the basin. Fluids of different composition moved laterally at different levels of the pile and repetition is related to successive inputs of basic magma high into the crust. This indicates that mineral assemblages may result from buffering by stratal waters, local high temperature gradients, rock composition, and metastable effects, all or some of which indicate there is no simple relation to depth.

M. P. ATHERTON

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Cross-references: *Greenschist facies*; *Metamorphic facies*; *Ocean-floor metamorphism*; *Prehnite-pumpellyite facies*; *Pumpellyite-actinolite schist facies*; *Zeolite facies*.

BUSHVELD COMPLEX

The *Bushveld Complex* is situated in the central Transvaal, South Africa, and covers an area of 67,000 km² of which about one-third is concealed by younger formations (Fig. 1). It consists of a large variety of extrusive, intrusive, and metamorphic rocks, the basic stratified portion of which constitutes the largest layered intrusion of its kind in the world. This suite of igneous rocks was emplaced in the following chronological order: (1) volcanic products preceding the emplacement of the hypabyssal and plutonic rocks; (2) basic sheets injected into the sedimentary country rocks; (3) the Rustenburg Layered Suite represented by the layered sequence of ultrabasic, basic, and intermediate rocks; and (4) a later plutonic phase represented by the Bushveld granite.

Volcanic episode

Willemse (1969) considered the andesitic lavas, tuff, and agglomerate in the Pretoria Group of the Transvaal Supergroup to be the first manifestation of the major igneous event. The volcanic activity culminated in the outpouring of large masses of rhyolite and

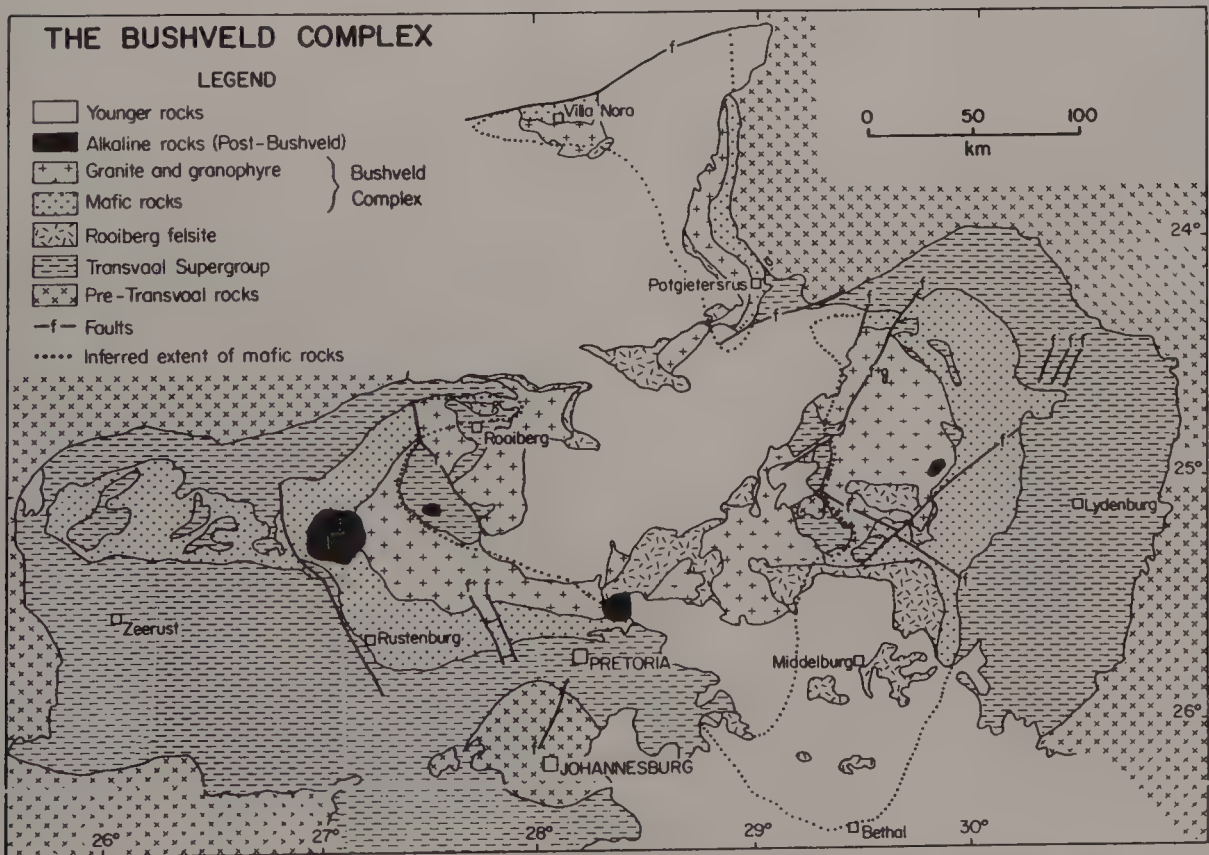


FIGURE 1. Generalized geological map of the Bushveld Complex. (After Willemse, 1969.)

associated agglomerate and tuff, collectively referred to as the Rooiberg felsite (Twist and French, 1984). This sequence of acidic rocks is close to 3,000 m thick and together with leptite and granophyre, constitutes the “roof” of the layered sequence.

Basic sills

There are two types of basic sills. The older suite was intruded prior to emplacement of the

Rustenburg Layered Suite and is now represented by amphibolitic rocks, whilst the younger sills are unaltered, syn-Bushveld in age and considered to represent offshoots of the magma in the Bushveld chamber. They range in composition from peridotite, through micro-pyroxenite to gabbro and are spatially and genetically related to the marginal zone of the Layered Suite. The sills have a cumulative thickness of about 2.6 km and increase in frequency upwards in the host sediments.

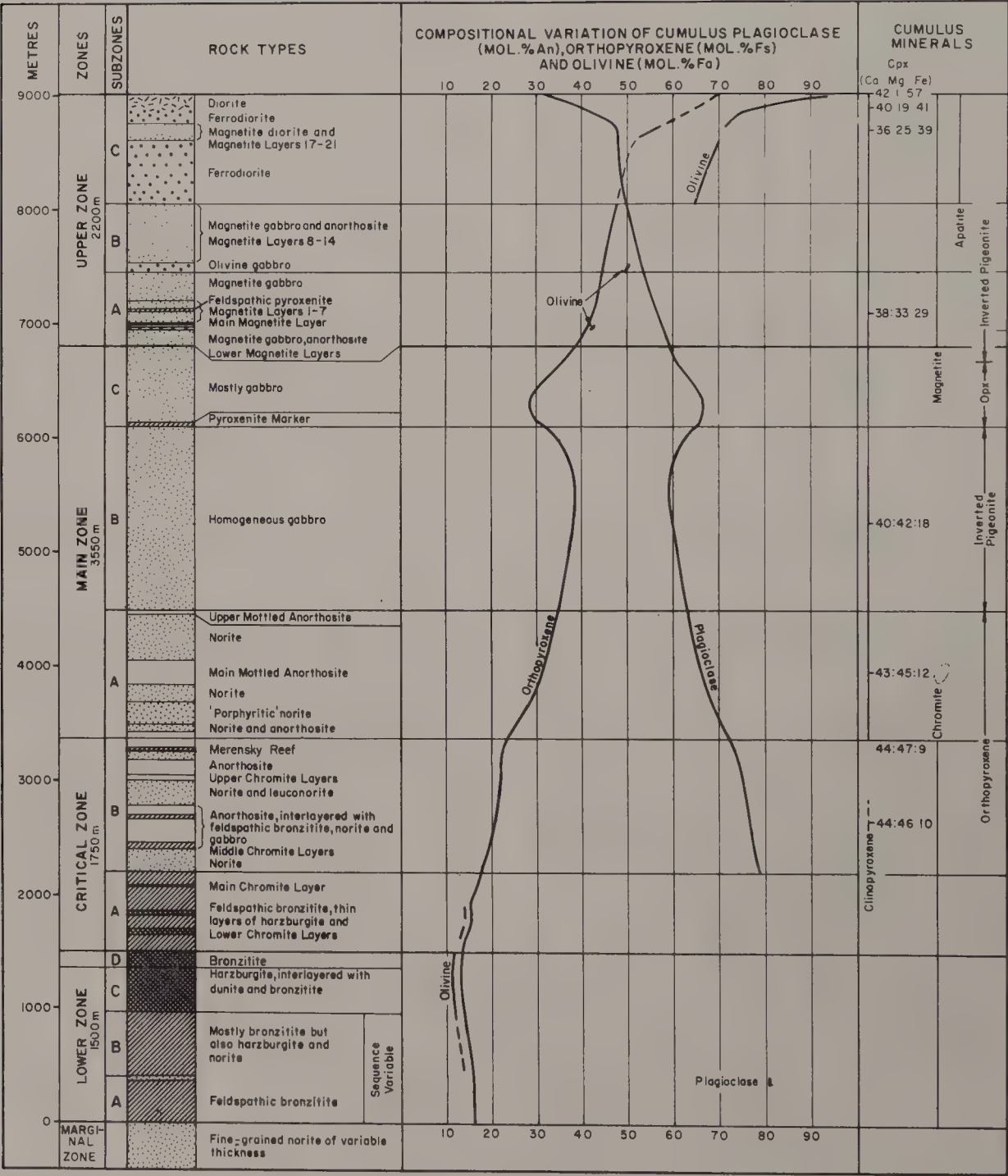


FIGURE 2. Subdivision and rock types of the layered sequence of the Bushveld Complex and compositional variation of the cumulus minerals.

Rustenburg Layered Suite (2050 ± 22 Ma)

This suite displays a remarkably complete sequence of the fractionation products of a tholeiitic magma in which almost all assemblages ranging from ultrabasic to acidic differentiates are developed (Fig. 2). These rocks occupy five adjoining basin-like masses, the outline of which at surface is essentially lobular (Fig. 1).

The intrusion was previously considered to be a lopolith (Hall, 1932) because of the parallelism of the layered rocks and the surrounding sediments. Willemse (1969), however, emphasized the transgressive nature of the layered rocks across marginally folded and faulted country rocks. This feature, in addition to gravity data, has led to the contention that emplacement of magma was from at least seven centers of eruption, resulting in several, overlapping, funnel-shaped intrusions, with crystallization taking place in separate, interconnected magma chambers (von Gruenewaldt, 1979).

Many of the more prominent layers such as the Merensky Reef and the main magnetite layer can be followed for more than 450 km along strike in the eastern and western lobes. This remarkable continuity of the individual layers, and the broad similarities of the sequences in the various intrusions, are perhaps the most striking features of this Complex (Fig. 3).

The following summarizes the great variety of cumulates and the processes responsible for their formation (see Wager and Brown (1968),

Cameron (1971), and Irvine et al. (1983) for further information).

Marginal zone. A variety of rocks varying in composition from pyroxenitic low in the succession to gabbroic at higher levels separate the rocks of the Layered Suite from the surrounding sediments. Similarity of these with the syn-Bushveld sills led to the recognition that two magma types were involved in the evolution of the Layered Suite (Irvine and Sharpe, 1982; Cawthorn and Davies, 1983; Sharpe, 1984). The earlier types are olivine boninitic in composition, rich in Cr and the Pt-group elements, whereas the later types are close to tholeiitic basalts. Mixing between these two magma types and density stratification of the derivatives within the magma chamber are considered important mechanisms for the origin of the layering and the extensive chromite and platinum-group element deposits (Sharpe and Irvine, 1983; Irvine et al., 1983).

Lower zone (basal zone). In the eastern and western lobes of the Complex, where the most complete successions of the layered sequence are developed, the rocks of the lower part of the lower zone occupy several adjoining troughs, separated from one another by upfolded rocks of the Transvaal Supergroup (Cameron, 1978). The sequence of rocks differs from one trough to the next, so that harzburgite may be the lowermost exposed type in one whereas it is feldspathic bronzitite in another. Over the whole area, the upper part of this zone is characterized by pronounced rhythmic layering, caused by a repetition of layers of harzburgite and bronzitite (Fig. 3). Chromite, although present as an accessory cumulus phase, forms layers of economic importance only in the Potgietersrus area.

Critical zone. The critical zone is a remarkably varied sequence of cumulates of either bronzite, chromite or plagioclase alone, or various combinations of these. The layers vary from uniform cumulates, a few hundred metres thick, to sequences showing rapid vertical changes and consisting of layers varying from less than 2 cm to several metres in thickness (Cameron, 1971). In the lower half, bronzitites are most abundant, whereas plagioclase-rich cumulates predominate in the upper half. Several chromitite layers occur throughout the sequence, the most important of which is the 1.5 m thick Steelpoort or main layer.

The upper part of the critical zone is characterized by a number of cyclic units, the chromitite layers of which are enriched in Pt. The best developed of these cyclic units is the Merensky unit with the Merensky Reef at its base (Vermaak, 1976). This reef, a feldspathic, coarse-grained pyroxenite with a very thin

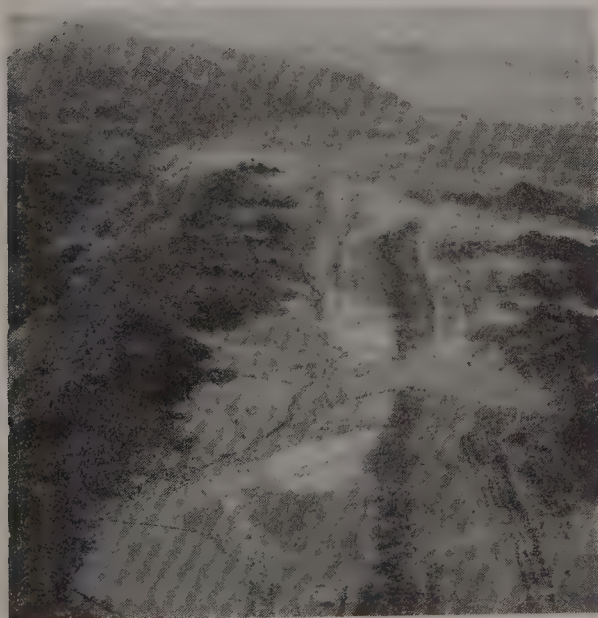


FIGURE 3. Layered pyroxenite of the lower zone in the eastern part of the Bushveld Complex, looking ESE; whitish soil is underlain by harzburgite.

chromitite layer at its base, is well known as an important Pt, Ni, and Cu deposit.

Main zone. The overlying, lower third, of the main zone consists of alternating layers of plagioclase-rich cumulates such as leuconorite and anorthosite. Gabbro-norite constitutes the majority of the remaining rocks of the main zone. They are very homogeneous in nature and are characterized by the presence of inverted pigeonite. Of interest is the marked compositional break at the base of the pyroxenite marker, which is considered to be due to an influx of large amounts of fresh undifferentiated magma into the chamber (Fig. 2).

Igneous lamination, a texture characteristic of most of the gabbro and norite, has its best development in this zone. The texture is due to a preferred orientation within the plane of layering of tabular, cumulus plagioclase grains, but in some instances also of needle-like crystals of pyroxene (Willemse, 1969).

Upper zone. The appearance of cumulus magnetite in the sequence marks the beginning of the upper zone. More than twenty magnetite layers, usually associated with plagioclase-rich cumulates (anorthosites) and varying in thickness from 50 mm to more than 10 m are developed. The main magnetite layer contains on an average about 1-6% V_2O_5 and forms the largest deposit of vanadiferous iron ore in the world. Fe-rich olivine and apatite also enter as important cumulus phases and their presence serves as a basis of further subdivision of this zone. Ferrodiorite is the characteristic rock type of subzone D, whereas rocks of intermediate composition only occupy the topmost 100 m of the intrusion. The most acidic differentiates, of granodioritic composition, occur as small, pocket-like intrusions in the overlying roof rocks (von Gruenewaldt, 1972).

Metamorphic effects

An extensive metamorphic aureole, several kilometers wide in places, is developed in the floor of the layered sequence. Sillimanite-bearing hornfels and granulite of the pyroxene hornfels facies are developed close to the contact of the intrusion, whereas cordierite, staurolite, chiastolite, and chloritoid are constituents of the hornfelses farther away. Metamorphic mineral assemblages indicate temperatures of up to 1,000°C at pressure of ca. 0-5 kb.

Metamorphism of the roof of the layered sequence resulted in the recrystallization of the Rooiberg felsite to leptyte and microgranite as well as the melting of these to produce a magma which crystallized as granophyre (von Gruene-

waldt, 1972). The felsite, leptyte and granophyre are sometimes collectively referred to as the epicrustal phase of the Bushveld Complex (Willemse, 1969).

Lebowa Granite Suite (1,920 ± 40 Ma)

The Nebo granite (Bushveld granite) occupies extensive areas in the central portion of the Complex, where it is intrusive into felsite, granophyre, and leptyte. The latter two usually separate the granite from the basic rocks, whereas felsite and granophyre constitute the roof. It can be regarded as a basin-shaped, sheet-like intrusion of batholithic dimensions. Over the larger part of the area the granite is a homogeneous, coarse-grained, pink to gray rock, and is considered to be responsible for the Sn and fluorspar mineralization within the confines of the Complex. Porphyritic and granophyric varieties are encountered along the margin of the intrusion. The Makhutso granite (1,670 ± 30 Ma) occurs as stocks in the Nebo granite.

GERHARD VON GRUENEWALDT

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C

CALC-ALKALINE ROCKS

Calc-alkaline rocks are a broad group of volcanic and plutonic rocks found in modern island arcs, cordilleran continental margins, and ancient orogenic belts. The group includes a range of compositions and mineralogies ranging from mafic basalts, through intermediate andesites to the felsic and silicic dacites and rhyolites. The critical definitions are chemical: Peacock (1931) used the term to describe rocks with total SiO_2 in the range 55–61 wt.%, where $\text{Na}_2\text{O} + \text{K}_2\text{O}$ wt. % / CaO wt. % was >1 . Such rocks contain little or no free silica (quartz, tridymite, cristobalite), are dominated by feldspars, hornblende and/or augite, and never contain feldspathoids, alkali pyroxenes, or alkali amphiboles.

A calc-alkaline "suite" is defined on the basis of a large geochemical data set; individual rocks may then be termed "calc-alkaline rocks," i.e., members of a "calc-alkaline trend." The exact limits and internal subdivisions of this suite are not precise, and some debate surrounds whether boninites (high-Mg, ultramafic andesites) and shoshonites (high-K andesites) should be included. On an AFM diagram the calc-alkaline suite displays a different trend from the tholeiitic suite by lacking Fe-enrichment, and displaying a rapid increase in SiO_2 content in the less-mafic members (Fig. 1). Qualitatively, the tholeiitic suite contains few intermediate members, often being bimodal, whereas in calc-alkaline suites the intermediate members are usually predominant volumetrically.

Volcanic calc-alkaline rocks are represented by the series high-Al basalt–andesite–dacite–rhyolite, with the plutonic equivalents being gabbro–tonalite–granodiorite–granite. The volcanic rocks, especially the andesites, dacites and rhyolites, tend to be porphyritic, while the basalts are often aphryitic and glassy. Typical phenocrysts include oscillatory-zoned plagioclase. Ca-rich and Ca-poor pyroxenes, usually augite and hypersthene rather than pigeonite, occur both as phenocrysts and groundmass components. Some andesites contain amphibole, and the more felsic rocks may contain biotite; both these minerals are more important in the plutonic members of the suite.

Examples of bulk chemical composition of

typical calc-alkaline rocks are given in Tables 1 and 2. A characteristic feature is the low and more or less constant FeO/MgO ratios, contrasting with the rapid Fe-enrichment relative to SiO_2 content in the tholeiitic suite. On the AFM plot this is expressed by marked alkali enrichment and minimal Fe enrichment versus SiO_2 content (Fig. 1). The Na/K ratios vary widely, reflected by variable modal plagioclase/alkali feldspar ratios.

Compared to their equivalents in other series, calc-alkaline rocks are rich in Al_2O_3 , some being peraluminous. In contrast the high-field-strength elements (HFSE) Ti, Zr, Y, and P are typically lower in abundance. The Al_2O_3 vs. SiO_2 contents of calc-alkaline rocks is antipathetic, with the mafic members being more Al_2O_3 -rich than the felsic-silicic rocks. Among the basalts, the alkali abundances are intermediate between alkali basalt and tholeiitic basalt. The most abundant members of the calc-alkaline suite

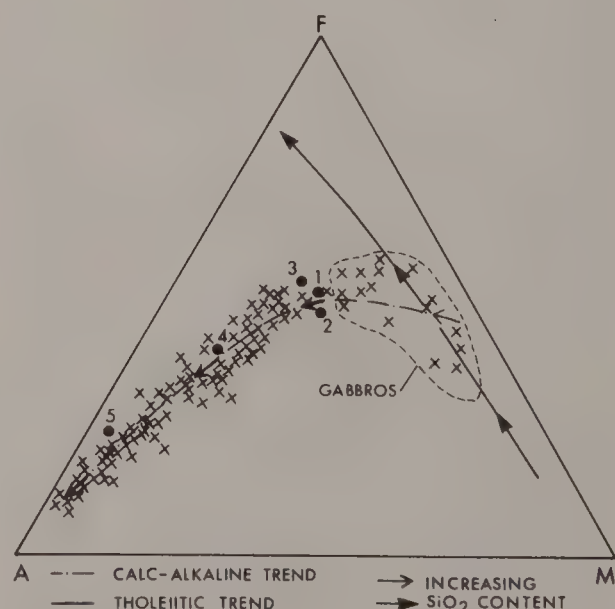


FIGURE 1. A (alkalies: $\text{Na}_2\text{O} + \text{K}_2\text{O}$)– $\text{F}(\text{Fe}_2\text{O}_3 + \text{FeO})$ – $\text{M}(\text{MgO})$ diagram showing the calc-alkaline and tholeiitic trends for igneous rock compositions; numbers correspond to average analyses in Table 1; crosses are analyzed calc-alkaline plutonic rocks from the Peruvian batholith. (After Atherton et al., in Atherton and Tarney, 1979.)

TABLE 1. Chemical compositions of calc-alkaline volcanic rocks

	1	2	3	4	5
SiO ₂	50.80	55.10	59.70	65.01	73.23
TiO ₂	1.27	0.82	0.70	0.58	0.24
Al ₂ O ₃	18.70	17.80	17.10	15.91	14.03
Fe ₂ O ₃	—	3.30	2.80	2.43	0.60
FeO	9.70 ^a	4.90	3.80	2.30	1.70
MnO	0.12	0.16	0.12	0.09	0.02
MgO	5.00	4.30	3.20	1.78	0.53
CaO	9.50	8.10	6.60	4.32	1.32
Na ₂ O	3.30	3.10	3.30	3.79	3.94
K ₂ O	1.44	1.20	1.50	2.17	4.08
P ₂ O ₅	0.32	0.21	0.19	0.15	0.05
H ₂ O	—	0.93	1.00	0.19	0.58
CO ₂	—	—	—	0.06	—
TOTAL	100.15	99.90	99.00	99.78	100.14

^aTotal Fe as FeO.

1. High-Al basalt (average of 17), Java and Bali; Whitford et al. (1979).
2. Basic andesites (average of 648); Gill (1981).
3. Acidic andesites (average of 888); Gill (1981).
4. Dacites (average of 578); Le Maitre (1976).
5. Rhyolites (average of 8), Cascade province, NW U.S.A.; Carmichael (1964).

(andesites) have bulk compositions similar to estimates of the terrestrial crust. Conversely, the important role of continental crust in the bulk composition of orogenic andesites has been shown by Francis et al. (1977) and Thorpe and Francis (1979) by Sr- and Nd-isotope studies.

TABLE 2. Chemical compositions of calc-alkaline plutonic rocks

	1	2	3	4	5
SiO ₂	48.26	57.48	61.52	66.09	72.08
TiO ₂	0.62	0.95	0.73	0.54	0.37
Al ₂ O ₃	19.07	16.67	16.48	15.73	13.86
Fe ₂ O ₃	2.78	2.50	1.83	1.38	0.86
FeO	6.98	4.92	3.82	2.73	1.67
MnO	0.18	0.12	0.08	0.08	0.06
MgO	7.96	3.71	2.80	1.74	0.52
CaO	11.20	6.58	5.42	3.83	1.33
Na ₂ O	1.97	3.54	3.63	3.75	3.08
K ₂ O	0.24	1.76	2.07	2.73	5.46
P ₂ O ₅	0.12	0.29	0.25	0.18	0.18
H ₂ O	—	1.36	1.24	1.04	0.53
CO ₂	—	0.10	0.14	0.08	—
TOTAL	99.41	99.98	100.01	99.90	100.00

1. Gabbro (average of 8), Peru; Atherton and Tarney (1979).
2. Diorite (average of 755); Le Maitre (1976).
3. Tonalite (average of 83); Le Maitre (1976).
4. Granodiorite (average of 723); Le Maitre (1976).
5. Calc-alkaline granite (average of 72); Nockolds (1954).

Within both island arcs and cordilleran belts the composition of calc-alkaline rocks (major elements, trace elements, and isotopes) shows consistent variations. These variations display spatial and secular trends orthogonal to the subduction zone, and may be related to the depth of the subducted oceanic slab and the thickness of underlying continental crust and the mantle wedge.

Plutonic rocks in the calc-alkaline suite follow the same general trends, with two main types (and intermediate varieties) being recognized. For granitic members these are referred to as being I- and S-type (see **Granite**), whose characteristics reflect the nature of their source regions (Chappell and White, 1974).

Genesis

Calc-alkaline rocks are the dominant magmatic suite in orogenic belts, so their origin is one of the most important problems in petrology and bears on the way in which continental lithosphere is generated. Most attention has focused on the origin of the andesites, and particularly on three fundamental questions. (1) Why are calc-alkaline rocks so intimately associated with destructive plate margins? (2) Why does their trend in Fe/Mg ratios differ from that of the tholeiitic suite? (3) Why do they show conspicuous enrichment in large-ion lithophile (LIL) elements? For a review of the subject and comprehensive references see Atherton and Tarney (1979), Gill (1981), and Thorpe (1982).

Partial melting of the subducted oceanic slab. There appears to be an unique association of calc-alkaline volcanism with subduction zones, suggesting that melting of the subducted slab contributes to the parental magma. As it sinks to mantle depths the subducted slab undergoes dehydration and metamorphism, first to amphibolite, then, at greater depths, to quartz eclogite. Melts of andesitic composition could be generated either by partial melting of quartz eclogite or amphibolite under high P_{H_2O} , at depths between 100 and 150 km. Such melts could rise directly to the surface as andesite or tonalite (Green and Ringwood, 1968). However, the major and trace element abundances in erupted andesites differ markedly from those predicted in these models. Kushiro (1973) suggested that partial melting of dehydrated ocean-floor basalt at 20 kb pressure (≈ 60 km depth) could produce high-Al basalt liquid, and that progressive further partial melting of the residuum at depths up to or greater than 100 km would produce progressively more silicic melts. Between 100 and 200 km depth such melts would resemble

rhyodacite lavas in composition, and such liquids may in turn instigate melting in the overlying mantle wedge.

Partial melting of mantle peridotite. Experimental studies have suggested the possibilities of producing andesitic melts directly from mantle peridotite under high P_{H_2O} . The interpretation of such results has been fraught with controversy (Gill, 1981), as experimental melts may not represent equilibrium assemblages, and they differ substantially from erupted andesites in their trace-element content, Fe/Mg ratios, and water content. Only the high-Mg andesites or boninites have the primitive characteristics predicted for this mode of origin, and though calc-alkaline they cannot be considered parental to the whole suite.

Enrichment processes in the mantle source. Even the most primitive, mafic calc-alkaline rocks show enrichment in LIL elements compared to either mid-ocean ridge basalts (MORB) and chondritic mantle standards. This suggests a LIL-enriched source region for the parental melt (Pearce, in Thorpe, 1982). The relative depletion in HFSE may be due to the stabilization of a Ti-bearing phase during melting, e.g., a kaersutitic amphibole, reflecting high P_{H_2O} or P_{O_2} in the source region. Two models may account for this: (1) introduction of LIL-enriched and silicic melts from the subducted slab into the mantle source region, or (2) metasomatism of the source region by LIL-bearing fluids released from the dehydrating slab (Gill, 1981; Pearce, in Thorpe, 1982).

Fractionation of a basaltic parent magma. Crystal fractionation in these circumstances must provide mechanisms that permit SiO_2 and LIL enrichment, HFSE depletion, and suppression of strong Fe-enrichment, in contrast to the processes and mechanisms pertaining to the tholeiitic series. Magnetite fractionation due to high P_{O_2} may account for some of this (see review by Kushiro, 1979), while the coprecipitation of plagioclase + orthopyroxene $\pm$ olivine + augite + magnetite under low pressures may explain most of the observed chemical and petrographic characteristics of these rocks (Gill, 1981), not least the most commonly observed intratelluric phenocrysts.

Green and Ringwood (1968) suggested amphibole fractionation as the prevailing mechanism (see Gill, 1981, for detailed reviews). But amphibole stability may be too restricted to account for the huge volumes of erupted andesite. Furthermore, amphibole phenocrysts are rare in erupted rocks of this suite, in contrast to their occurrence in some intrusive complexes (Irvine, 1974). There remains the possibility that amphibole fractionation or crystal fractionation in general, is

important in the evolution of plutonic calc-alkaline rocks, but is less important, or insignificant, in the petrogenesis of the related volcanic suite.

Contamination and mixing of basaltic magma with crustal material. The "shoshonite trend" of K_2O enrichment away from a subduction zone suggests that the thickness of continental crust may relate to calc-alkaline rock composition. This could reflect contamination of an evolving parent magma, or mixing of primary basalt with progressively more crustal melt away from the subduction zone. Thermal constraints in that hybridization and melting of crustal rocks are heat-consuming processes, will not allow such a mechanism to operate in isolation. However, in concert with crystal fractionation, such assimilation may account for many features of calc-alkaline rocks.

Crustal anatexis. Subdivision of granites into I- and S- types (Chappell and White, 1974) presume a large crustal contribution to calc-alkaline plutonic rocks with the melting involved requiring a crustal thickness in excess of 35 km. Although this process may account for the more silicic plutonic rocks, and generate the parents of rhyodacite and ignimbritic volcanic rocks, a mantle component appears necessary to account for most high-Al basalts and andesites. Furthermore, it is difficult to envisage a crustal source for magmas like andesite and tonalite whose liquidus temperatures exceed 1,000°C with 2% H_2O .

Integrated petrogenetic models. One petrogenetic model alone is unlikely to satisfy all the demands presented by such a complex suite of rocks; given the complexity of interacting processes at destructive plate margins, this is hardly surprising. Furthermore, because destructive plate margins themselves display considerable heterogeneity, and as more evidence for heterogeneity in the mantle source region is obtained, it may be unrealistic to expect one model or one synergistic mechanism to be ubiquitous. Each of the mechanisms reviewed here may operate somewhere and at some time, but the precise role and importance of each one may vary from locality to locality.

J. SOUČEK, F. V. HOLUB, E. JELÍNEK,
H. KLÁPOVÁ,

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Cross-references: *Andesite line*; *Basalt*; *Circum-Pacific plutonic and volcanic activity*; *Granite*; *Igneous rocks—tectonic setting*; *Island arcs—petrology*; *Pacific and Atlantic suites*; *Volcanic rocks*.

CALDERAS AND VOLCANO-TECTONIC DEPRESSIONS

Calderas are large-scale kettle- or cirque-like depressions found in areas of volcanic activity. They are destructional in origin and most have resulted from the collapse of a portion of the crust overlying a magma reservoir. In plan view they range from circular to elliptical to U-shaped; in diameter they range from ca. 1 km to ca. 25 km. The following are examples of

calderas.

	Diameter (km)
Bandaisan	2.5
Mt Somma	3.3
Kilauea	3.3 × 4.2
Okmok	8.2
Crater Lake	10.0
Aniakchak	10.0 × 10.8
Asosan	16.5 × 25
Valle Grande	26.5 × 30

Circular to elliptical calderas are typically formed by collapse; they are concentric or nearly so with the volcanoes on which they form. Their shape in plan view is probably determined by the shape of the underlying lava reservoir or by the similar plan of the volcanoes on which they form. This latter factor determines the stresses that would result if support were withdrawn from below.

Calderas that are U-shaped in plan view result from either eccentric collapse or from explosive eruption. The bottom of the U is usually located directly below the summit region of the original volcano.

Another large and important group of caldera-like depressions have typically rectilinear boundaries, which are a reflection of some regional tectonic pattern. Such depressions, *volcano-tectonic depressions*, are distinguished from calderas proper not only by their appearance in plan view but also by their generally much greater areas. The following are examples:

	Dimensions (km)
Lake Ranau	12 × 17
Taupo–Rotorua graben	25 × 100
Lake Toba	30 × 100
Yellowstone Plateau	42 × 75

Calderas differ from craters in a number of ways, but the principal distinction is in origin. The walls of a crater are built up by the eruption of lava or pyroclastic material onto the surface, while the walls of a caldera are produced either by the blowing away of a portion of a volcano or, more commonly, by its collapse into the underlying lava reservoir. A crater is thus a primary feature of the volcano, a caldera is a secondary one.

Craters and calderas also differ in size. The area of the floor of a crater, determined by the cross-section of the conduit leading from the lava reservoir to the surface, is rather small. The floor area of a caldera is related to the diameter of the upper part of the lava reservoir itself and so may be very large. Some overlap in size between craters and calderas is to be expected; however, probably few if any craters exceed a floor diameter of 1–2 km.

Theories of origin of calderas

The origin of calderas has long been a thorny problem. The following summarizes the more acceptable theories (see Williams, 1941).

By explosion. According to this theory, large calderas are similar in origin to the smaller craters, and the difference in size is merely a measure of the intensity of the explosions that produce them. In support of this view, the great volumes of material erupted within historic times by several volcanoes is often cited. However, the extensive sheets of pyroclastic material surrounding most calderas consist almost exclusively of fresh magma, and the proportion of old lava fragments is inadequate to account for the disappearance of the tops of the former cones.

By gas-coring and insliding. Escher (1933) has stated that if a large cylinder is drilled by explosion, insliding of the walls may take place on so great a scale as to form depressions many kilometers in width. Subsequent rise of lava may then convert the funnel-shaped depressions into flat-floored calderas. From experiments with artificial volcanoes, he concluded that the radius of the eruption cylinder is a function of the velocity of the eroding gas; to produce a large cylinder, the lava reservoir must be both very deep and very large, and the amount of gas must be enormous. In this case, extremely large volumes of rock fragments must be blown out prior to caldera formation, and no such volume of lithic ejecta can be found near any caldera. Also, the forms of caldera floors unmodified by post-caldera activity are unlike those postulated by the theory.

By internal solution. Assuming that a volcano encloses a large reservoir of liquid lava, this might grow larger as a consequence of melting of the walls. Should the magma remain inside the volcano, it would slowly crystallize to form a resistant core. However, if the opening of lateral vents drained much of the magma from the interior of the cone, the top of the solid shell might collapse to produce a caldera. Alternately, a caldera might be formed by a solution of the walls of a crater by a lava lake within it. There is no doubt that magmas as different as rhyolite and basalt can effect solution of older lavas under near-surface conditions; however, the persistence of narrow and closely-spaced vents in many craters suggests that the widening of conduits by solution is an abnormal process. Much more light might be thrown on the probable extent of these processes by detailed studies of the pyroclastic deposits surrounding calderas, but the lack of such studies allows no definite conclusions to be drawn as to the importance of these processes.

By collapse. By far the most important cause of caldera formation seems to be the removal of magmatic support, although collapse may also occur by ring-fracture stoping where the load of the volcanic edifice becomes too great. The principal way in which support is withdrawn is by the rapid evisceration of the magma chamber by the copious outpouring of fluid (basaltic) lava from fissures far down on the flanks of the volcano or by the violent and rapid explosion of enormous volumes of pyroclastic material, dominantly in the form of acidic pumice. To a lesser extent, evisceration may occur by the intrusion of magma into suddenly-formed fissures to produce dykes and sills.

Calderas formed by explosion

Calderas formed by explosion appear to be restricted to the larger extrusion domes. They are generally of small diameter and are U-shaped in plan view. The rock that originally filled the caldera was broken up and ejected laterally from the open end of the U, probably in the form of an avalanche, to produce an apron- to stream-like deposit of rock fragments that were, for the most part, derived from the dome itself.

In some cases (e.g., Bandaisan in northern Honshu in 1888) the explosion that produced the caldera was apparently a phreatic one. Such explosions, which may occur long after growth of the dome has ceased, take place when meteoric water penetrates into the still hot core of the dome. The avalanches then produced are generally of relatively low temperature, and there is usually no post-caldera eruptive activity.

In other cases the explosion may be truly volcanic (e.g., Bezmyanny in 1956), occurring during the growth of the dome. The avalanches that result are composed of incandescent gas-evolving fragments; the evolution of these gases greatly increases the mobility of the avalanches to the point where they may be more appropriately called agglomerate flows. These flows represent one extreme of the collapse type of peléan eruption (see **Pyroclastic eruptions**). In this case, formation of the caldera may be followed by a renewal of lava extrusion to build up a new dome on the floor of the caldera.

Collapse calderas

Calderas produced by collapse may be formed in a number of ways. The collapse itself usually follows partial evacuation of the lava reservoir underlying the volcano, though in some cases it may be due simply to overloading of the crust above the lava reservoir. Most collapse calderas can be assigned to one or another of four types.

Calderas of the *Krakatoa-type* (Figs. 1, 2) are

essentially confined to stratovolcanoes and so are most common in orogenic zones such as the circum-Pacific region, although examples are also known in the E African rift system (Leat, 1984). The collapse is typically piecemeal, following the massive eruption of pyroclastic material; the eruption is usually of the plinian type and the ejected material is composed largely of acidic pumice, but other types are known (Self et al., 1984). The flanks of the remnant of the cone are usually blanketed with this material, of which only 5–10% is composed of fragments of old lava. Post-caldera eruptions may occur almost anywhere on the floor of the caldera. These may result in the building up of one or more cinder cones, as at Crater Lake, Oregon, U.S.A. On the other hand, a whole new stratovolcano may be built within the caldera, as at Vesuvius, until the only evidence for the caldera is an arcuate ridge (or somma) that partly encloses the new cone, and even this may be buried eventually.

Calderas of the *Glen Coe*-type result from the process of cauldron subsidence. A roughly cylindrical plug, bounded by ring fractures that are vertical or dip steeply outward, sinks into the lava reservoir (Fig. 3); this sinking may result from evisceration of the lava reservoir, but it may also result simply from overloading

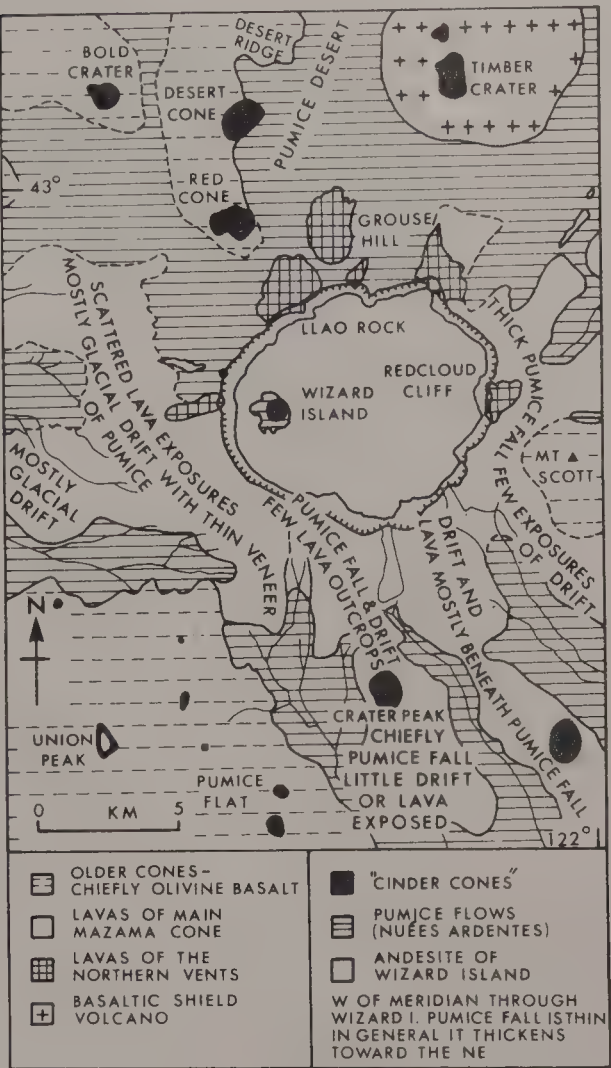


FIGURE 2. Map of Crater Lake caldera, Oregon, U.S.A. (After Williams, 1941.)

of the roof of the reservoir. Magma rises up the open space between the plug and the country rock; it may solidify before reaching the surface, to form a ring dyke, or it may be erupted onto the surface, to build up a series of post-caldera cones around the margin of the caldera. The

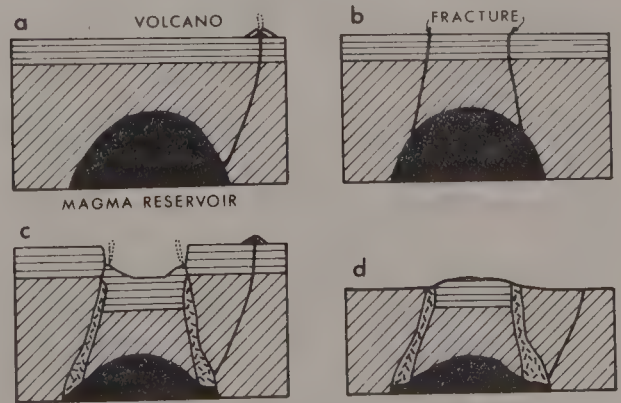


FIGURE 3. Stages in the development of a Glen Coe-type caldera. (After Billings, 1942.)

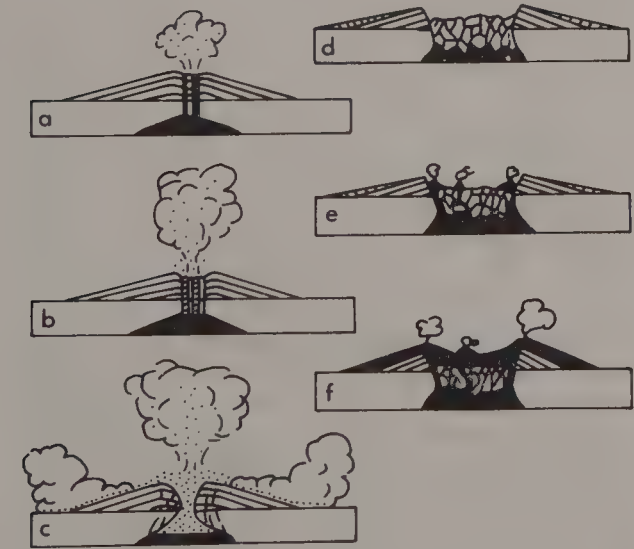


FIGURE 1. Evolution of a caldera of Krakatoa-type. (After Williams, 1941.) (a) Mild explosions of pumice; magma stands high in conduits. (b) Explosions increase in violence; magma level falls in the main chamber. (c) Culminating explosions; part of the ejecta is hurled high above the cone but more of it rushes down the flanks as nuées ardentes; magma level deep in chamber; roof begins to crack. (d) Lacking support, the top of the cone collapses into the magma chamber. (e) After an interval of quiescence and erosion, new cones appear on the caldera floor, especially near the rim. (f) Young cones rise above the caldera rim and pour lavas down the outer slope.

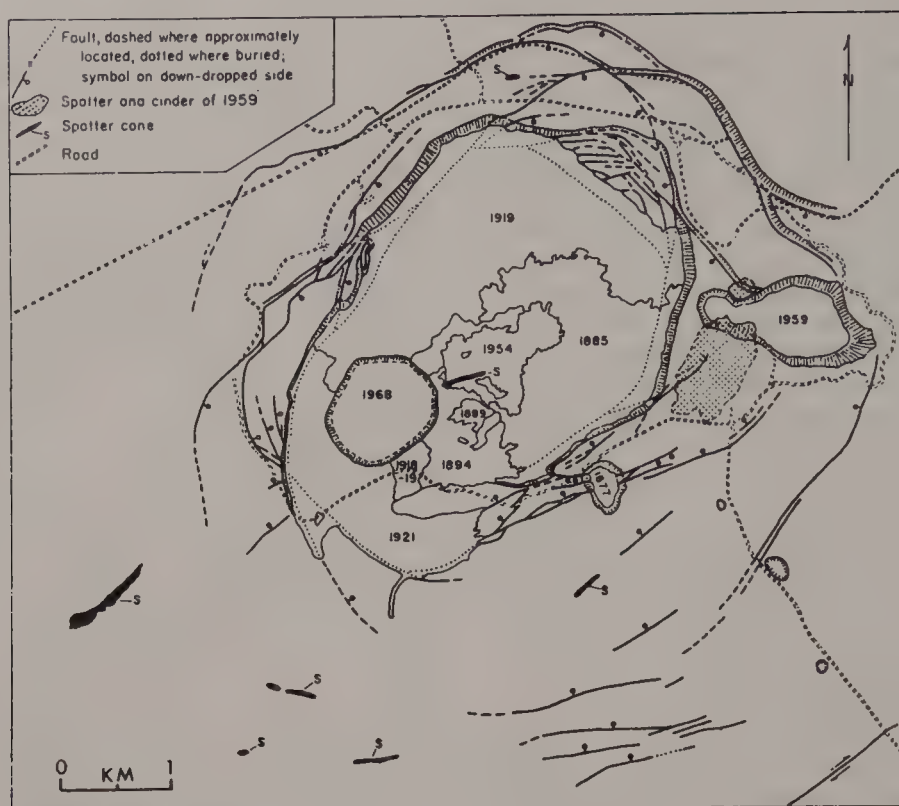


FIGURE 4. Kilauea crater, Hawaii. (From Macdonald and Abbott, 1970.)

development of such a ring of vents will build up a wall around the caldera and so accentuate its appearance. Examples of this type of caldera include Medicine Lake in northern California, U.S.A., and Lake Coatepeque in El Salvador.

Calderas of the *Kilauea-type* appear to favor shield volcanoes; the best examples are Mokuaweoweo and Kilauea, Hawaii (Fig. 4). Collapse occurs along steep inward-dipping ring fractures. The conical block so defined may result from the down-dropping of a keystone-like block when a period of strong tumescence is followed by rapid withdrawal of lava from the reservoir. This withdrawal is usually a result of the copious effusion of lava from vents far down on the flanks of the volcano. Alternatively, it may result from the gradual enlargement of a steep-walled summit pit crater by slumping of the sides and engulfment of the debris during periods of low stand of magma in the conduit. The structural terraces that occur on the walls of some Kilauea-type calderas may represent such slump blocks. Calderas of this type probably develop gradually over a period of time. Post-caldera activity is the rule; in most cases the lavas erupted into the caldera eventually fill it to the brim. This post-caldera activity occurs from a pit crater within the caldera (Kilauea) or from one or more fissures in the floor (Mokuaweoweo).

Calderas of the *Valles-type* are restricted to ignimbrite shields, the broad low dome-like

edifices built by the eruption of voluminous pyroclastic flows from central vents (Fig. 5; see Smith and Ross, 1961). Prior to the emission of these pyroclastic flows, the basement (usually composed of earlier volcanic rocks) is domed up by the rise of a gas-rich relatively acidic magma. In this dome, cone fractures develop that allow the vesiculation of the magma and serve as conduits for the eruption onto the surface of the pyroclastic flows that build the ignimbrite shield. Simultaneously, the crest of the dome subsides along these cone fractures to form a caldera, the floor of which will be covered with the same ignimbrite that forms the flanks of the shield. If the now gas-depleted magma continues to rise, the floor of the caldera also will be up-arched to form a resurgent dome: the caldera is referred to as a *resurgent cauldron*. The new cone fractures developed within it, or the older ones at its margins, may act as conduits to produce a ring of lava extrusions within the caldera and surrounding the resurgent dome (Aldiss and Ghazali, 1984).

Volcano-tectonic depressions

These are essentially basins, bounded by more or less straight fault scarps, where the sinking of the central block is a result of volcanic processes or of a combination of volcanic and tectonic processes. In many cases they occur in the crestal portions of major

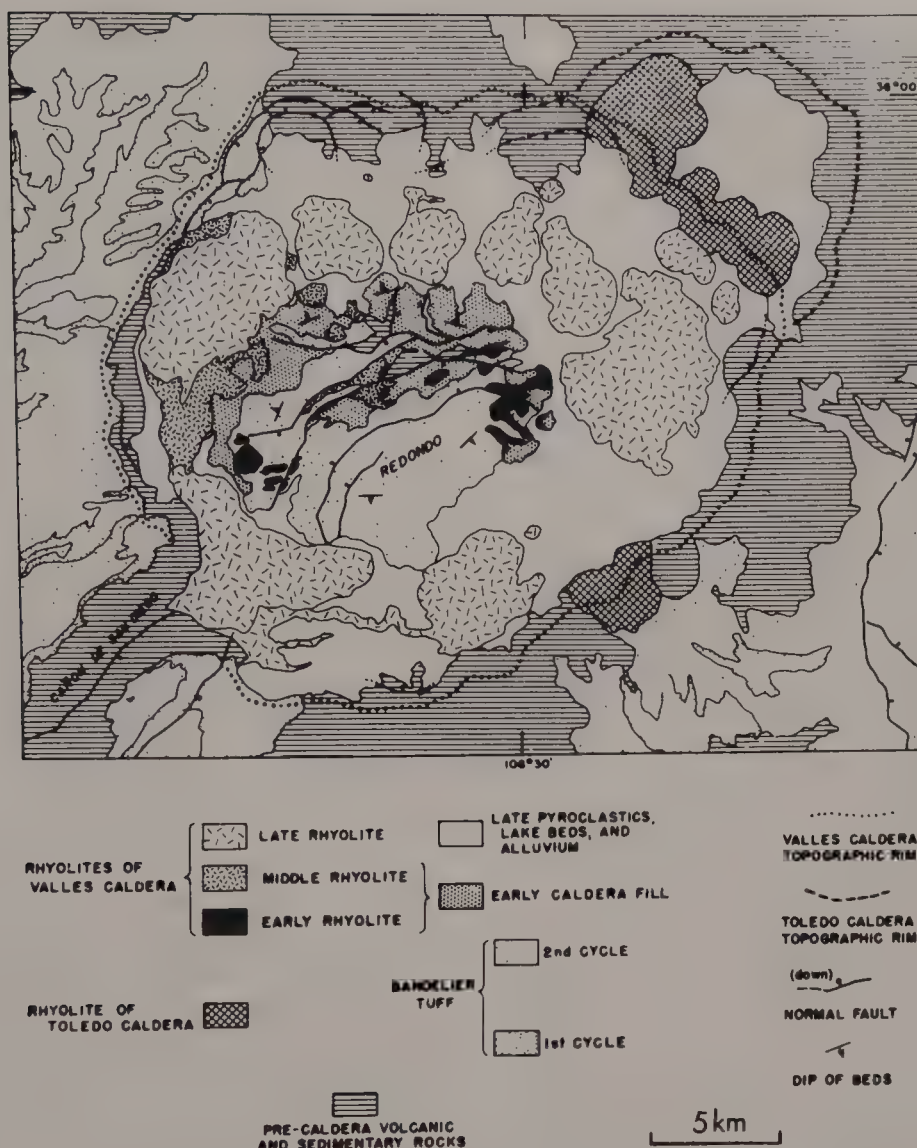


FIGURE 5. Valles Caldera, New Mexico, U.S.A. (From Smith and Bailey, 1968.)

anticlines. They differ from calderas in that their outlines are controlled by regional rather than by local lines of weakness, but all gradations exist between them and Valles-type calderas. In all cases the volcanic process involved seems to be the emission of enormous volumes of usually acidic pyroclastic material in eruption of rotoruan, or rotoruan and plinian types; this results, in turn, in the collapse of the crust. In many cases the depression formed is itself filled partly or completely by lava and pyroclastic material; the total volume of the material erupted may fall in the range of 800–2,000 km³.

CHARLES P. THORNTON

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Cross-references: *Explosive volcanic eruptions—classification; Granites and volcanism; Nuée ardente; Pyroclastic eruptions; Ring dyke; Volcanoes and volcanism.*

CARBONATITE

Carbonatite is an igneous rock, found as subvolcanic intrusions and as lavas and tuffs, usually consisting largely of calcite and/or dolomite. There are also alkalic (or natro-) carbonatites, which consist of complex carbonates of Na, Ca, and K. All are now considered to be essentially of magmatic origin, although sometimes modified by metasomatism. Most occurrences involve three related rock suites: (1) carbonatites, (2) alkaline and/or ultrabasic eruptive rocks, and (3) country rocks affected by intense K- and/or Na-metasomatism (*fenites*); the whole association constitutes a carbonatite complex. The associated alkaline rocks are varied, usually nepheline-bearing types, especially ijolites, or nephelinites, frequently melilitic, but in general miassic rather than agpaite types. The ultrabasic types

include olivinites and diopsidic pyroxenites, and in certain cases a kimberlite–carbonatite relationship is evident, although in several regions (e.g., part of Ontario, Canada, and Zambia) typical carbonatites appear to have no associated alkaline rocks. Excluded, however, are rocks replaced by carbonates in other circumstances, such as carbonated serpentinites and spilites, metamorphic rocks such as sagvandites, and metamorphic limestones even where associated with alkaline rocks, as in the limestone–nepheline gneiss terrane of the Grenville province of Canada.

The classic occurrences include Fen and Alnö, Scandinavia, Iron Hill, Colorado and Magnet Cove, Arkansas, U.S.A., and the Chilwa province, Malawi, first described, respectively, by W. C. Brögger, H. von Eckermann, E. S. Larsen, F. Dixey, and W. Campbell Smith. Since 1950 hundreds of new examples have been found, many of great petrologic and/or economic interest, and the numbers still grow. Complexes in many countries are reviewed by Tuttle and Gittins (1966), Heinrich (1966), Currie (1976), and Le Bas (1977, 1987).

Size, structure and form

Intrusive carbonatites. Carbonatite complexes cover 1–250 km² in area, of which carbonatite may account for between 1% and >70%. The area of carbonatite is usually small, 2–3 km² ranking as large. However, the Sokli complex in Finland has a 5 km² core of magmatic carbonatite, set in a 15 km² ring of largely metasomatic carbonatite, surrounded by 30 km² of fenites. At Sukulu, Uganda, carbonatite occupies 18 km². The classic complexes are circular or oval in plan, carbonatite usually occupying a central core or plug, or forming arcuate intrusions, either ring dykes or systems of cone sheets (Fig. 1). These circular structures are interpreted as subvolcanic diatremes, drilled usually by a succession of intrusions, three or four of which may be discernible at present erosion levels (Fig. 2). Sometimes only carbonatites and fenites are exposed, or two related intrusions may occur only a few kilometers apart, one largely of alkaline rocks, the other largely carbonatite, as exemplified by the Jombo and Mrima intrusions in Kenya, and Tchivira and Bongo in SW Angola. Other carbonatites form dyke-like intrusions up to 20 km in length emplaced along fault planes, sometimes recurrently over distances of up to 100 km (e.g., Karema depression and Songwe scarp in Tanzania). What appears to be a large layered sill of carbonatite agglomerate occurs in Mesozoic sandstones at Kaluwe in Zambia, not

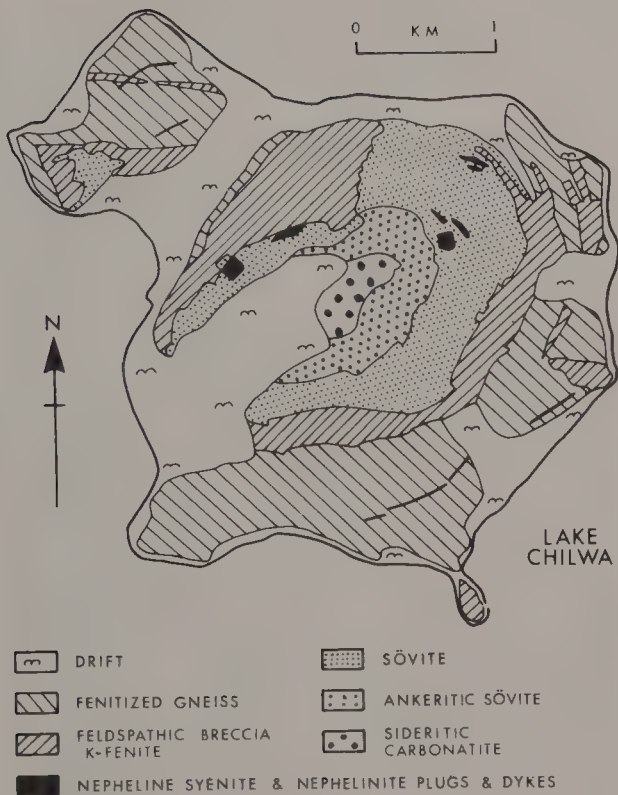


FIGURE 1. Simplified geological map of Chilwa Island carbonatite complex, Malawi. (After Garson, 1962.)

far from a group of carbonatite volcanoes. This “sill” is 100–240 m thick, and crops out over 17 km².

Extrusive (volcanic) carbonatites. Dissected volcanoes revealing an intrusive plug of carbonatite in a pile of feldspathoidal lavas and tuffs, usually nephelinitic (e.g., Napak, Uganda and Kaiserstuhl, W Germany) first suggested the volcanic setting of carbonatites. Other examples among the Neogene volcanoes of E Africa are Elgon and Kisingiri, both formerly 100 km in diameter. Indications of extrusive carbonatites were confirmed in 1960 when eruptions of molten carbonatite lava were witnessed and sampled in the crater of the volcano Oldoinyo Lengai in N Tanzania. This volcano is built largely of nephelinite and phonolite tuffs and lavas, and, unexpectedly, the carbonatite proved to be anhydrous alkalic carbonatite. This rock undergoes incongruous dissolution by water, and is rapidly converted to

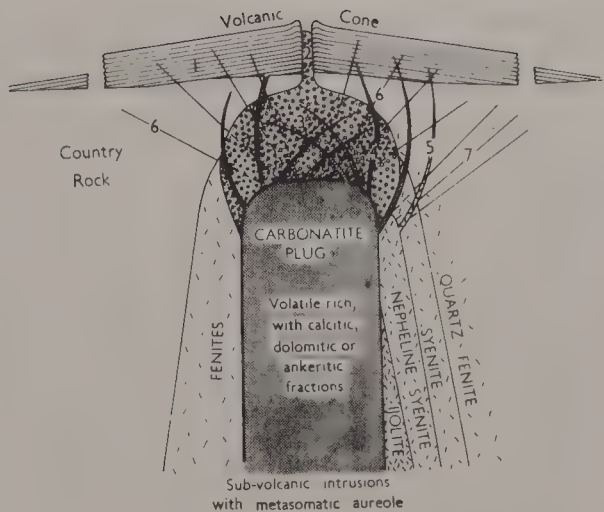


FIGURE 2. Schematic representation of a carbonatite complex; diameter of plutonic part of complex 1–10 km and of volcanic cone 15–60 km: 1, tuffs and lavas, mainly nephelinite, often carbonated; 2, vent filled with agglomerate and carbonate; 3, zone of brecciation—disrupted K-fenite and carbonatite—with recrystallization and carbonate replacement; 4, 5, ring dykes of carbonatite (4) and nepheline syenite (5); 6, 7, carbonatite cone sheets from final focus of eruption—cone sheets from earlier focus of eruption also occur. (Modified after Garson, 1962.)

calcite, so that in older volcanic fields its former presence is only indicated by calcified replacements. Oldoinyo Lengai remains unique, but older extrusive carbonatites are now known in many countries, both in Africa and elsewhere, e.g., Afghanistan (Khanneshin, a well-preserved Quaternary volcano), and Greenland (Qagssiarssuk, of Proterozoic age).

Composition and nomenclature

Intrusive carbonatites. These vary in composition and texture. Table 1 sets out five types with their principal carbonate, grain size, geological occurrence, and type areas. This nomenclature is not ideal, and in 1980–83 various new schemes were proposed (Kresten, 1983, citing earlier papers) but some points are still debated. Woolley (1982) suggested a threefold chemical classification comprising

TABLE 1. Composition and nomenclature of intrusive carbonatites

<i>Sövite</i>	Calcite	} Medium- to coarse-grained, in plugs and dykes; Fen, Norway.
<i>Rauhaugite</i>	Dolomite	
<i>Alvikite</i>	Calcite	} Fine-grained, in narrow dykes and cone sheets; Alnö, Sweden.
<i>Beforsite</i>	Dolomite	
<i>Ferrocarbonatite</i>	Calcite (Fe-rich)	Fine-grained, dark late-stage dykes; W Kenya.

calcic carbonatites (sövite, alvikite) with $\text{CaO} > 80\%$ of $\text{CaO} + \text{MgO} + \text{FeO} + \text{Fe}_2\text{O}_3 + \text{MnO}$, *magnesiocarbonatites* (rauhaugite, beforsite) with $\text{MgO} > \text{FeO} + \text{Fe}_2\text{O}_3 + \text{MnO}$, and *ferrocarbonatites* (Le Bas 1977) with $\text{FeO} + \text{Fe}_2\text{O}_3 + \text{MnO} > \text{MgO}$. The accessory silicates include feldspar, biotite, aegirine, arfvedsonite, and melanite garnet but total nonsilicates, especially apatite, magnetite, sulfides, zircon, and traces of pyrochlore and baddeleyite, may be as abundant, or more so. When the carbonate is $<70\%$ (by volume), silicate-rich types may be termed *silicocarbonatites*, or the dominant accessory may prefix the name, e.g., magnetite sövite, apatite beforsite. However, silicified carbonatites, i.e., those replaced by late-stage hydrothermal silica, should not be termed silicocarbonatites.

Enrichment in F, P, Nb, Sr, Ba, and REE is a distinctive geochemical feature, and readily distinguishes carbonatites from limestones and marbles. Carbonates have usually recrystallized, with loss of original primary igneous textures, and developed granoblastic fabrics. In many places early intrusions of sövite with magnetite, apatite and pyrochlore are followed by dolomitic or ankeritic carbonatites, containing Mn, REE, Sr, and Ba. The latter include *rare-earth carbonatites* such as the dolomite-barite-bastnaesite rock mined at Mountain Pass, California, U.S.A., and the ankerite-strontianite-monazite rock of Kangankunde, Malawi. These particular phases suggest that crystallization temperatures may have ranged from perhaps 600°C down to as low as 100°C .

Extrusive carbonatites. These comprise the unique, unstable lava of Oldoinyo Lengai (called *lengaitite*), pyroclastic rocks of the same composition, and mixed nephelinite-lengaitite tuffs, and also, at older centers, calcified alkalic carbonatite pyroclastic assemblages. The solubility of lengaitite has contributed largely to the high Na-content of the nearby lakes, Natron and Magadi. Lengaitite consists mainly of nyerereite and gregoryite (complex Na-Ca-K carbonates), respectively, forming characteristic lath or barrel-shaped crystals, and calcite pseudomorphs after nyerereite typify the alkali-free calcified alkalic carbonatite pyroclastic rocks of areas such as the Tinderet foothills, Kenya (Deans and Roberts, 1984), where clasts of trachytoidal calcified nyerereite lavas, with or without phenocrysts of primary calcite, are common.

The pyroclastic rocks include lapilli, Pele's tear-drop lapilli, and accretionary lapilli tuffs; alkalic carbonatite tuffs were ideal media for the rapid fossilization of terrestrial faunas (including primates), and the preservation of spectacular "footprint beds" (at Laetoli, Tanzania).

Because of these fossils, the carbonatite matrices were for many years accepted as lacustrine limestones.

Carbonatite-fenite association

The marked contrast in composition between lengaitite and the deep-seated carbonatites, which are alkali-free, is explained by the universal association of fenites with intrusive carbonatites. These alkalic metasomatic rocks indicate that the primary carbonatite magma was alkaline (like lengaitite, but perhaps not of identical composition), but very unstable, and on cooling released all the alkalis to fenitize the wall-rocks, leaving only calcemic carbonatites. The common fenite is usually a greenish sodi-potassic rock with alkaline pyriboles, not unlike the fenites associated with alkaline syenites and ijolites. It may be accompanied, however, by various almost monomineralic rocks: (1) ultrapotassic feldspar rocks with 12–15% K_2O (potash fenites), commonly associated with large masses of high-level sövite; (2) albitites, which are much rarer; (3) phlogopite-biotite rocks or xenoliths (glimmerites) that appear in narrow zones around dolomitic carbonatites and (4) magnesio-arfvedsonite rocks (blue fenites) found in shears and fault zones at some distance from the carbonatite. In ultrabasic-carbonatite complexes, other products of metasomatism are found, one widespread type being an apatite-magnetite-serpentine (or forsterite) rock (*phoscorite*), which develops between carbonatite and diopsidic pyroxenites.

Conditions of emplacement

Carbonatite plugs, dykes, and cone sheets, emplaced in some cases at depths down to 8 km, appear to have crystallized at temperatures ranging from ca. $800\text{--}600^\circ\text{C}$ (upper limit) to perhaps 400° or 300°C in later dolomitic differentiates. Gas pressure (mainly CO_2) was very high, and explosive disruption of solid carbonatite and wall-rock is often evident in tuff-dykes, vent agglomerates and carbonatite breccias, probably emplaced as fluidized (gas-solid) systems. In volcanic eruptions the great preponderance of nephelinitic and carbonatitic tuffs points to highly explosive conditions, despite the known high fluidity of lengaitite magma. Metasomatic fronts probably advanced ahead of the uprising volatile-rich magma, and may have been replenished as crystallization progressed. In the final stages, alkaline and saline hydrothermal solutions were frequently active, probably involving mobilized ground-water, and sometimes depositing quartz, fluorite, or barite. Carbonatite activity thus extends

from low-temperature magmatic to postmagmatic stages.

Geographical distribution and geological setting

Carbonatites are worldwide, being known in at least 130 districts in some 44 countries (Fig. 3). They appear to be absent from the volcanic girdle of the Pacific Ocean, and from the volcanic facies of most older destructive plate margins, being instead characteristic of the stable interiors of continental plates, especially where deep crustal fractures are inferred. They are abundant and best exposed in the region of the rift zones of E Africa, where they are of late Precambrian, Mesozoic, Tertiary, and Recent ages. Many occur near major tectonic intersections, such as the junctions of rifts, but often the locations are less easily explained. Another prolific area is the Canadian Shield, where, although exposures may be poor, geophysical surveys have located them in unexpected numbers. Rectilinear chains or belts of carbonatites are not solely related to lines of rifting, and often coincide spatially with lines of major basic and alkaline complexes of differing

ages. Thus, the carbonatites of Transvaal, S Africa, lie in a broad latitudinal belt through the axis of the Bushveld Complex and the Pilanesberg alkaline massif; the carbonatites of Mozambique and Malawi form an E–W belt running through the much older gabbro–anorthosite complex of Tete; and in southern Ontario and Quebec a line of carbonatites of various ages runs through the Sudbury norite complex and is continued westwards by the Monteregian alkaline intrusions. However, there are several apparent exceptions to these generalizations. In the E Atlantic Ocean carbonatites occur in the Cape Verde and Canary Islands, and isolated examples are known in fold mountain ranges, including the Rockies, Andes, High Atlas, and the Southern Alps of New Zealand.

Economic value

A surprising variety of mineral deposits occurs in carbonatites, but important mines are limited to some 15 complexes, and their combined outputs for the year 1983 (in thousand tonnes, and decreasing order of sales value) reveal the main products involved:



FIGURE 3. World distribution of carbonatites; each triangle represents either a single complex or a district with two or more complexes: additional complexes occur at Sarfartoq, Greenland (66°30'N, 52°51'W), Qaqarsuk, Greenland (65°25'N, 51°40'W), Bear Lodge Mtns., Montana, U.S.A. (44°35'N, 104°20'W), Cerro Impacto, Bolivar, Venezuela (6°0'N, 65°50'W), and Tajno, Poland (53°45'N, 23°15'E).

apatite 7,000, copper 134, pyrochlore 23, bastnaesite 23, vermiculite 200, magnetite 1,000 (?), baddeleyite 12, fluorite 21. Including byproducts, the total estimated value in 1983 was approximately U.S. \$700 million. About half of this total comes from three open-pit workings in the unique Palabora complex in Transvaal, S Africa. Here apatite (2,800) is in the peripheral zone of the carbonatite plug and adjacent phoscorite, and also in the metasomatized areas of the surrounding pyroxenite. The unique copper (134, and many byproducts) orebody forms a vertical pipe in the core of the carbonatite plug; the vermiculite mine lies in the weathered zone of an oval area of coarse diopside-vermiculite-olivine rock. Brazil is an equally important apatite producer (3,100 from four mines), and there are smaller operations at Siilinjärvi, Finland (381), Kovdor (Kola), U.S.S.R., and in Zimbabwe. The larger Brazilian apatite mines work secondary residual weathered deposits, but lower-grade primary deposits are also economic both in Brazil and Finland. From 1950 to 1966 Nb (pyrochlore) and rare-earth elements were the main products of carbonatite mines, and although soon eclipsed by Cu, and then by apatite, they have continued to meet the bulk of the western world's requirements of these elements. At

Araxa, Brazil, the largest Nb mine in the world works a rich residual pyrochlore deposit (near large residual apatite workings), and smaller pyrochlore mines operate at Catalao, Brazil, and at Chicoutimi, Canada. Of the several known rare-earth carbonatites, only the bastnaesite-barite carbonatite at Mountain Pass, California, U.S.A., is mined, and is the largest source of rare-earth elements in the world (see the survey by Deans, 1980).

Theories of origin

Interpretations of a magmatic origin of carbonatites met widespread criticism for many years because there was no evidence for the liquidus temperatures of about 1,300°C that were thought necessary. It seemed more plausible to invoke limestone assimilation or hydrothermal replacement. Since 1950, however, the mass of new field and petrographic data, coupled with experimental studies by P. J. Wyllie and others, have substantiated the magmatic hypothesis. Both hydrous calcium carbonate and Ca-alkali carbonate melts are stable to below 600°C at subvolcanic pressures.

The only natural carbonate melt observed to date was the alkali carbonatite lava erupted at Oldoinyo Lengai (Dawson, 1962). Eruptive

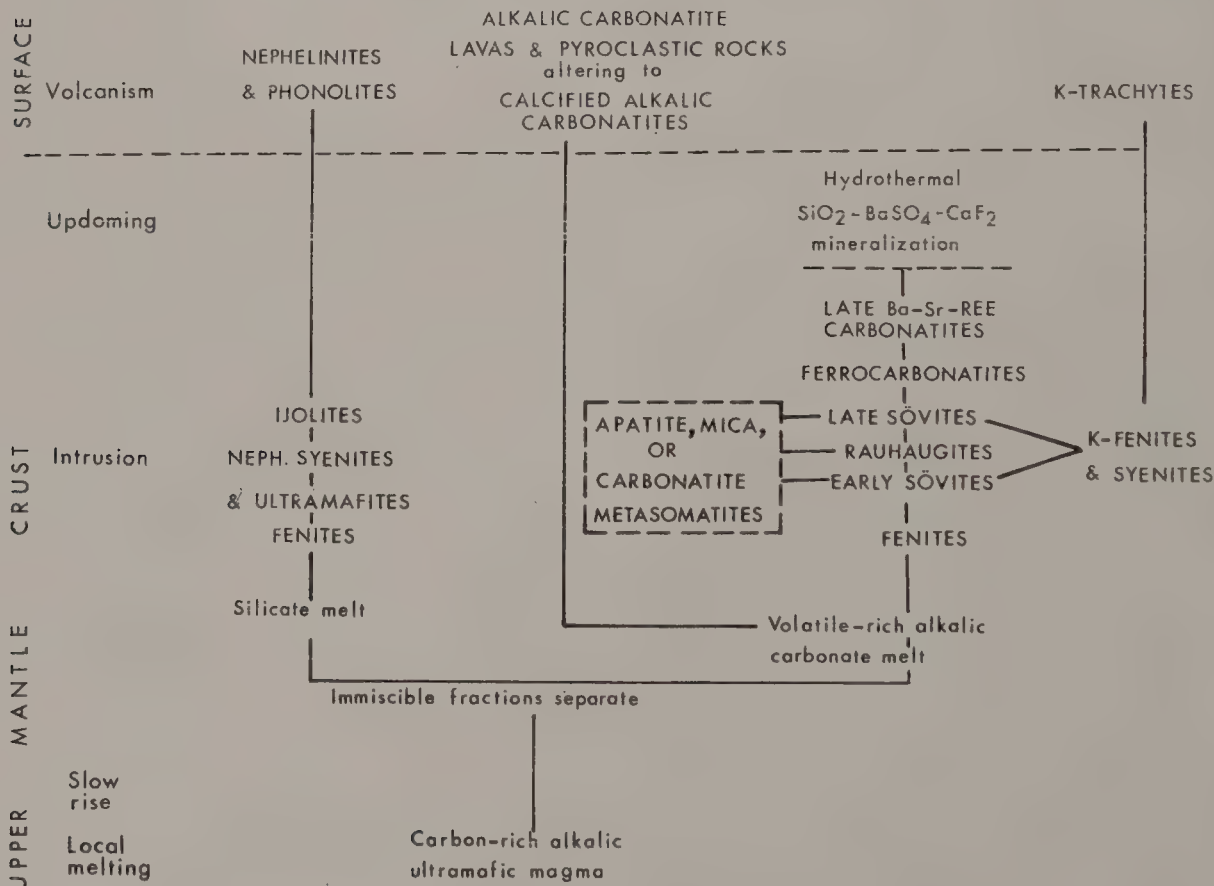


FIGURE 4. Petrogenetic interpretation of the evolution of carbonatites and related alkali metasomatic rocks.

carbonatite rocks with strong textural similarities have recently been described (Deans and Roberts, 1984), ranging back to the Proterozoic in age. Deeper-seated, intrusive carbonatites however, contain little or no Na and K and so the problem of the relationship between the eruptive alkali carbonatites and the intrusive alkali-deficient carbonatites is posed. Several lines of evidence indicate that presently alkali-deficient intrusive carbonatite bodies were formerly of alkali carbonatite composition. (1) The fluid inclusions in apatite crystals within alkali-deficient carbonatites approximate closely in composition to the alkali carbonatite lava of Oldoinyo Lengai (Rankin, 1975; Le Bas, 1981), indicating the presence of an alkali carbonate liquid during apatite crystallization. (2) The well-documented extensive fenitization adjacent to alkali-deficient carbonatite intrusions indicates the passage of abundant alkalies from the site of the intrusions into the country rock. It seems likely therefore that the presently alkali-deficient carbonatite bodies lost alkalies during emplacement and crystallization and that the composition of the original melt was comparable to that of the Oldoinyo Lengai lava.

The central problem of carbonatite origins thus becomes the origin of alkali carbonatite liquids. Igneous silicate rocks associated with carbonatite complexes consist of nephelinitic lavas and their plutonic equivalents and differentiates. The silicate lavas have geochemical characteristics indicating a mantle origin, whilst at the same time the Sr, Nd, and Pb isotope ratios of carbonatites are closely similar to those of the silicate lavas. This constitutes strong evidence of a common origin for both groups in the mantle. Given that liquid of carbonated nephelinite composition developed in the upper mantle by partial melting, there is experimental evidence to indicate (Freestone and Hamilton, 1980) that alkali carbonatite liquid could separate immiscibly from the parent to leave a nephelinitic liquid and thus become available for emplacement to yield either a carbonatite body with attendant fenitization, or alkali carbonatite extrusive rocks (Fig. 4).

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CATACLASTIC ROCKS

Cataclastic rocks are produced by *dynamic (cataclastic, dislocation) metamorphism* in discrete zones of differential stress and movement (process is termed *cataclasis* from Greek *kata*, down and *klasis*, a breaking), e.g., faults, thrust zones, and shear zones. Like all the products of metamorphism under stress (i.e., contact metamorphic rocks excluded) the features seen in the rock as exposed on the surface relate to the way the rock has reacted to stress and the manner of its recovery. Whether a rock reacts to stress by ductile or brittle behavior depends on its viscosity (competence), the pressure and temperature under which the stress acts, and the strain rate. The nature of recovery, i.e., recrystallization after or during the accommodation of that stress by strain, depends on the pertaining pressure and temperature regime. Brittle behavior is that which is seen where a

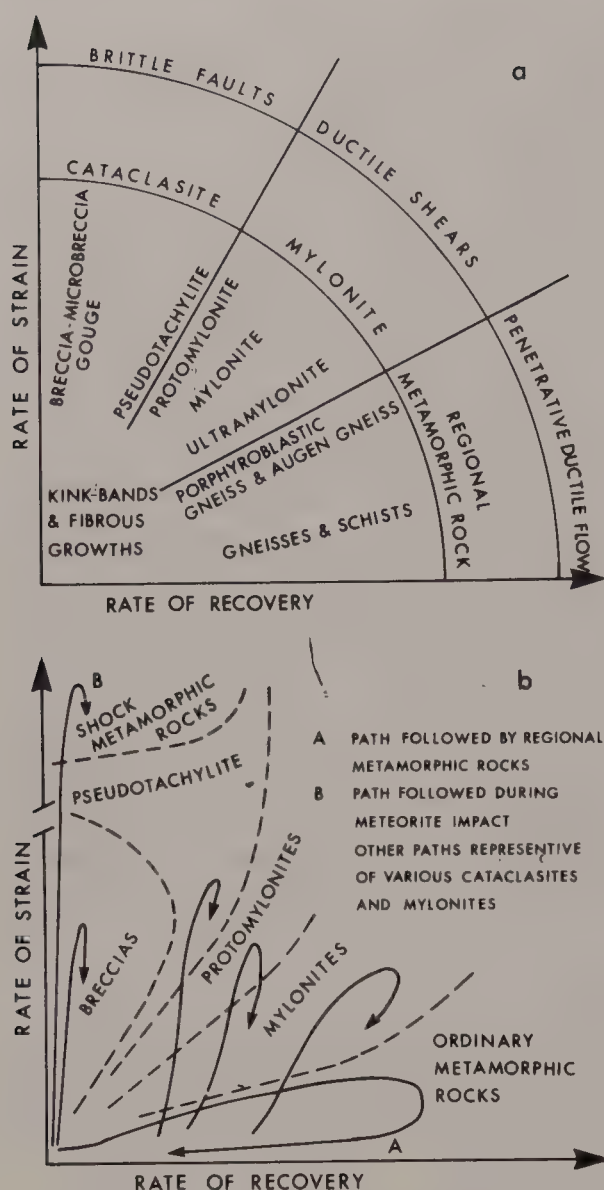


FIGURE 1. Schematic representation of (a) nomenclature and (b) conditions of formation of cataclastic rocks. (After Wise et al., 1984.)

rock fractures when strain is $< \text{ca. } 5\%$. Ductile behavior is that which occurs in a rock that deforms in a plastic fashion when strain exceeds, or greatly exceeds, $5\text{--}10\%$. Thus, under certain conditions a rock may undergo ductile, plastic deformation before suffering brittle failure, and brittle effects may be superimposed on ductile features. Strain may be accommodated by brittle fracture, ductile grain adjustments or pervasive ductile flow. Recovery is the process whereby crystallization processes anneal surplus grain-boundary or crystal-lattice free energy out of the system. Thus, at one extreme, shock metamorphic features are produced by enormous strain rates followed by minimal recovery, while regional metamorphic tectonites are the products of penetrative ductile

flow with optimal recovery (Fig. 1). The bulk of cataclastic rocks are the products of the interaction of strain rate and recovery intermediate between these two extremes. Shock metamorphism is considered separately (see **Shock metamorphism**).

The terminology applied to cataclastic rocks has been the subject of considerable confusion, particularly over the relationship of mylonite to other cataclasites. Lapworth (1885) introduced the term "mylonite" for flinty rocks in the Moine thrust zone (NW Scotland), which he regarded as indurated, extremely finely "milled" material. Mylonite was thus proposed to be the end-product of cataclasis, and hence an ultracataclasite. More modern concepts see mylonite as the product of ductile processes, whilst cataclasites are the products of brittle processes (see Wise et al., 1984).

Brittle phenomena associated with faults

These are high-level features. Brecciation is the predominant process, and recovery is minimal (Fig. 1a). The resulting rock-type is usually only poorly cohesive or even non-cohesive. These rocks contain many "stressed" features that have not been annealed by recrystallization. The rocks produced are structureless and are grouped collectively as *cataclasites*.

The distinction between *breccia* and *microbreccia* is one of grain-size, with $1\text{ mm--}0.5\text{ cm}$, depending on author, being the divide. Whereas breccia is dominated by rock fragments, microbreccia is often dominated by crystal or grain fragments. Cataclasites are usually poorly-sorted rocks, with microbreccia forming the matrix to larger clasts.

Frictional heating effects are minimal, but fluid action may give rise to a variety of alteration products like chlorite, serpentine, white mica, and clay minerals. Voids may be present and either empty, lined with drusy crystals, or be completely filled with carbonates, quartz or ore minerals. Consolidated, poorly-sorted debris in a fault zone, rich in clay, is referred to as *fault gouge*, or if unconsolidated as *rock flour*. Pervasive induration by micro-crystalline silica produces *silicified fault breccia*.

Irregularities in the attitude of the fault plane produces areas of compression and dilation. Breccia is largely confined to dilational zones (Fig. 2), and where the fault walls slide directly over each other. The polished surfaces are termed *slickensides*, which are often grooved by *slickenlines*. Fibrous growths may be present on these surfaces or filling cracks and voids, referred to as *slickenfibers* (Fleuty, 1975). A rock in which recognizable rock fragments



FIGURE 2. Schematic representation of features of cataclastic rocks in fault planes.

dominate (i.e., 50% by volume) is termed *protocataclasite*, and where the fine matrix forms >90% of the assemblage, the rock is termed *ultracataclasite*. There are also locally applied names (e.g., *kakirite* for a cataclastic rock at Lake Kakir, Swedish Lapland).

Pseudotachylite (pseudotachylite) is a cryptocrystalline (or glassy), dark, flinty rock, superficially resembling basic volcanic glass (tachylite). Some of this may be truly quenched melt, generated by frictional heating, and some show zoned spherulites and radially arranged microlites (Park, 1961). However, X-ray diffraction studies often reveal the presence of crystalline structures. It often fills fractures in an intrusive mode, and may represent ultrafine microbreccia, perhaps fluidized by “explosive” brecciation. Devitrification is common particularly in old pseudotachylite.

Cataclastic rocks are usually restricted to narrow fault zones, although many of these may run parallel or subparallel to each other across considerable areas, e.g., major continental wrench fault zones like the Great Glen fault zone in Scotland, or the Thompson or Nelson River lineaments in Canada; others cross the ocean floor. Cataclasis on a far larger scale occurs in the ocean trenches, where oceanic crust is subducted beneath island arcs or cordilleran continental margins. An accretionary prism of sediment and volcanic material scraped from the ocean floor may accumulate on the leading edge of the overriding plate, including within it chaotic zones where blocks of lava, gabbro, serpentinite, and pelagic sediment, often ranging from meters to kilometers in size, lie in a matrix of crushed and deformed pelagic shale. These are called *tectonic mélange* or *trench-wall megabreccia*. Such *mélange* assemblages occur throughout the Alpine-Himalayan belt, the western seaboard of the United States of America, and many locations around the Pacific Ocean.

Ductile features

Deformation is taken up in ductile behavior by movement between and within grains.

Increasing strain is marked by decreasing grain size, with strain energy being stored in lattice dislocations and deformations (seen as undulose extinction in thin section), and deformational twinning of crystals. Strain energy is released by reduction in grain size; this increases grain surface area. Recovery is marked by the annealing of strain grains and grain growth (decreasing grain surface area). The resultant rock is a product of the opposing interaction of these two processes.

Grain-size reduction begins around grain margins and along internal dislocations, producing a *mortar texture* of strained porphyroclasts in a matrix of finer-grained material. The whole grain assemblage is sutured throughout deformation, so that these rocks differ from cataclasites in always being cohesive. Where fragments of protolith constitute more than 50% of the rock, it is termed *protomylonite*; where porphyroclasts form less than 10% the rock is termed *ultramylonite*. Intermediate rock-types are termed *mylonite*, although Wise et al. (1984) recommend the use of *orthomylonite* here, using “mylonite” for the whole group, analogous to the use of “cataclasite.”

Porphyroclasts derived from phenocrysts or porphyroblasts, or in a lithology where layers underwent disruption by boudinage, undergo reduction in various ways. The former often retain a lozenge-shaped cross-section, or eye-like outline, forming an *augen* rock (German *auge*, eye). In other cases the porphyroclasts reduce to fine-grained aggregates that retain compositional integrity while being streaked out as *schlieren* or *flaser* fabric (e.g., *flaser gabbro*, *flaser gneiss*). Such features may give mylonite a layered or banded structure, complicated by intrafolial and drag-folds.

Quartz aggregates may recrystallize as oblate forms, with their *c*-axes characteristically perpendicular to the plane defined by the plate. In cross-section this platy quartz is termed *ribbon quartz* and may display undulose extinction along its length (cf., *plättung* texture in *granulite*, q.v.). Isolated mica flakes define a strong oriented fabric, and even slightly elongate quartz grains may share this. Large porphyroblasts may be fractured and disrupted, and frequently they are associated with *pressure shadows*, areas of unstressed crystal aggregates or fibrous grain growth (Spry, 1969). Mylonites thus differ from cataclasites in having a well-defined and often complex structure and pronounced grain fabric.

Recrystallization may proceed beyond annealing to include grain growth; the fine-grained mylonitic fabric coarsens. Where this process affects both the fine matrix and any porphyroblasts, the resulting rock is termed a



FIGURE 3. Blastomylonite with porphyroclasts of perthite (P) and a crystallized matrix of quartz (Q), albite (A) and biotite (B); XN, $\times 4$.

blastomylonite (Fig. 3). The term *hartscheifer* refers to a blastomylonite where only the matrix shows evidence of grain growth (Tyrrell, 1926).

Some mylonites may be brecciated, or fractured and “intruded” by veins of pseudotachylite. Such a relationship may reflect transport of the mylonite out of the deep ductile zone into higher levels where brittle processes occur. It may also represent high strain rates, where a rock initially deformed in a plastic fashion before undergoing brittle failure (Sibson, 1977). As vertical movement in fault zones may permit the juxtaposition of rocks from widely different depths, such superimposition of brittle features on ductile ones is not uncommon.

Fluid activity in mylonite brings about metamorphic and metasomatic as well as structural changes. In dry conditions, mineral alteration will include the inversion of orthoclase to microcline, but where water is introduced many more such changes are possible: e.g., plagioclase to epidote minerals, alkali feldspar to phengite and muscovite, and mafic minerals to talc, chlorite, and serpentine. Such rocks become rich in phyllosilicates like talc, mica, and chlorite, and become strongly

schistose. Such tectonic schists are termed *phyllonite*.

Cataclasite, mylonite, and fault mechanisms

Sibson (1977) relates the nature of fault rock to both depth and the nature of strain accommodation—whether stress is released by aseismic gliding or seismic brittle failure. The cataclasites result from an elastic and frictional mechanism, while mylonites are produced by plastic or quasiplastic deformation. In the case of the Outer Isles thrust, a major low-angle reverse fault belt in the Outer Hebrides, Scotland, Sibson relates the distribution of rock-types to palaeotemperature and the juxtaposition of rock masses from different depth zones (Fig. 4). A mylonite series (protomylonite–mylonite–ultramylonite) represents a ductile response to stress at depth, where stress is accommodated by continuous creep or gliding. The rock series protocataclasite–cataclasite–ultracataclasite $\pm$ pseudotachylite represents a brittle response to stress, where movement occurred incrementally following brittle failure and periodic stress release. This

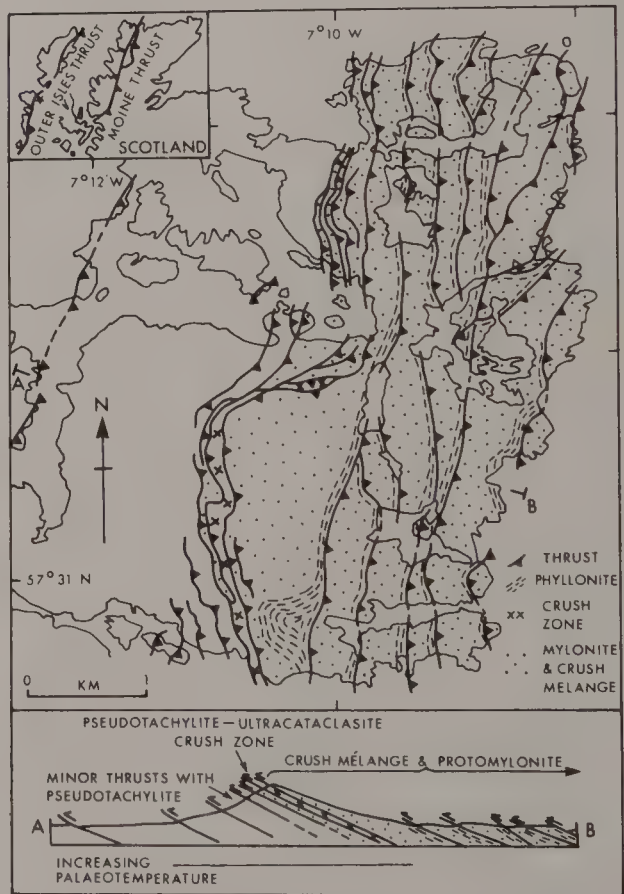


FIGURE 4. Distribution and relationships of mylonites, cataclasites, and thrusts, N Uist, NW Scotland (★) on inset. (After Sibson, 1977.)

brittle deformation is superimposed on mylonitic rocks as they were exhumed in the hanging wall of the fault zone.

An attempt to unify the nomenclature of fault rocks with that of fault mechanisms is included in Sibson (1977) and Wise et al. (1984).

Other terms

Using prefixes *cata-*, *hemi-*, and *holo-* from the Greek words *kata* (down, throughout), *hemi* (half), and *holo* (whole), and the suffixes *-clasis*, *-clastite*, from *klastos* (broken), Zeck (1974) sets out the following terms for types of cataclasite.

Hemiclastite—a cataclasite that shows only effects of weaker grades of cataclasis (*hemiclastis*); the nature of the original rock is still apparent.

Holoclastite—a cataclasite that shows advanced stages of cataclasis (*holoclastis*); the nature of the original rock is in doubt.

Protomylonite—a hemiclastite with a cataclastic foliation.

Mylonite—a holoclastite with a cataclastic foliation.

Ultramylonite—a mylonite in which the degradation is of a markedly advanced grade.

Pseudotachylite—a holoclastite consisting essentially of glass.

The following terms are also used by Zeck (1974).

Myloblastite—rocks that show the effects of ruptural deformation and a noteworthy degree of roughly simultaneous recrystallization (also hemi-myloblastites and holo-myloblastites).

Blastocataclasite—a cataclasite that has been recrystallized in part in a later, separate stage (also blastohemiclastite, blastohomoclastite, blastomylonite, blasto-ultramylonite).

ADRIAN F. PARK

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- Cross-references: *Gneiss*; *Granulite*; *Porphyroblast*; *Porphyroclast*; *Schist*; *Shock metamorphism*.

CHARNOCKITE

Charnockite is a felsic, melanocratic, holocrystalline, alkali granitic rock first described from the tombstone of Job Charnock in the graveyard of St. John's garrison church, Calcutta, India (Holland, 1900). The rock-type is widespread in the Madras area (whence the gravestone originated—St. Thomas Mtn.) and is known from much of the high-grade terrane of Peninsular India. Holland defined the rock as being of "granitic" composition with alkali feldspar and quartz that are typically clear but dark colored; the rock itself has a greasy appearance. The other essential phase in the assemblage is a pleochroic pyroxene, usually hypersthene, with a rhombic shape. The feldspathic component is dominated by K-feldspar, with albite or oligoclase as minor constituents. Garnet, biotite, iron-titanium oxides, sphene, apatite, zircon, and corundum are nonessential but frequent accessories. Myrmekitic and strained textures are very common in the Indian type examples (Fig. 1).

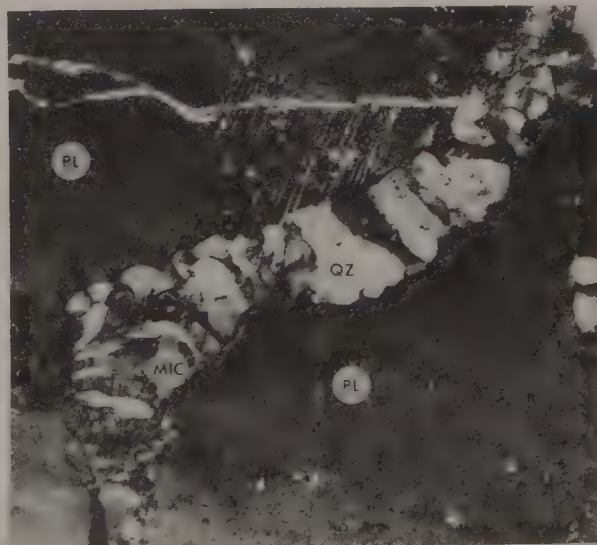


FIGURE 1. Microvein of myrmekitic charnockite neosome (quartz QZ, microcline MIC) between grains of strained plagioclase (PL) in basic granulite, Angul, Orissa, India. (From Park and Dash, 1984.) XN, $\times 9$.

In Peninsular India, charnockite is usually one of several neosomes in a migmatite complex that has undergone recrystallization under granulite or high-amphibolite facies conditions. There may be a spatial association with the incoming of granulite facies mineral assemblages, but the relationship of charnockite formation with this metamorphic process is controversial (cf., Janardhan et al., 1982; Powell, 1983; Raith et al., 1983). The charnockite is developed as patches at the expense of its host by a process of "charnockitization." At least in parts it was formed relatively late in a polyphase deformational sequence (Fig. 2—Park and Dash, 1984) and controlled by pre-existing structures, including conjugate sets of fractures (see Srikantappa et al. (1985) for color photographs).

Fluid-inclusion studies indicate the coexistence of the charnockitic neosome melt with a CO₂-rich fluid and have led to the suggestion that fluxing of hot CO₂-rich fluids of deep-seated origin resulted in the dehydration of amphibolite facies material to form granulite facies mineral assemblages: the carbon isotope composition of fluid inclusions in charnockites from southern India indicates a mantle source for the CO₂ (Jackson et al., 1988). "Charnockitization" induced by an isothermal decrease in P_{FLUID} relative to $P_{\text{LITHOSTATIC}}$ due to migration of the pore fluid into a network of fractures and ascent into higher crustal levels is advocated by Srikantappa et al. (1985).

In India the charnockite neosome has a fairly uniform composition (quartz + alkali feldspar) but its host can be a great variety of palaeosome lithologies. If the resultant variety of migmatitic rocks is considered to be charnockitic rather

than just a charnockite neosome in a variable palaeosome, it leads to the erroneous concept of a suite of pleochroic orthopyroxene-bearing rocks with the type charnockite as a hypersthene granite in this "charnockite suite." While the concept appears adequate to describe some deep-level, high-grade igneous-metamorphic complexes in other parts of the world, it does not apply to those of Peninsular India. The Indian type charnockites, a migmatitic neosome of regional extent in granulite or amphibolite terrane, appear to be distinct from the plutonic charnockite suite. Outside India there are similar occurrences in Sri Lanka, the Wheat Belt of W Australia, Enderby Land, Antarctica and the Lake Baikal region, Siberia, U.S.S.R., with a wide range of ages and evidence to point towards development in the deeper levels of a rifting environment that had a relatively high heat flow (cf. Aftalion et al., 1988).

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Cross-references: *Amphibolite facies*; *Charnockite suite*; *Granite*; *Granulite facies*; *Migmatite—origin of neosome*; *Migmatite—structural relationships*.

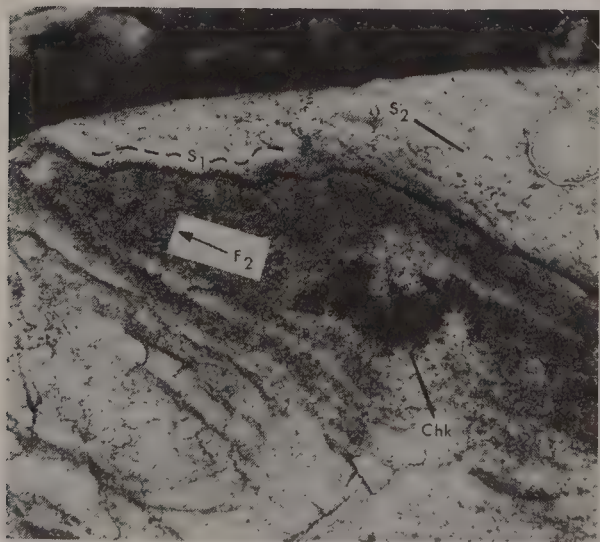


FIGURE 2. Patch of dark charnockite neosome (Chk) developed in folded migmatitic gneiss, Badakera, Orissa, India. (From Park and Dash, 1984.)

CHARNOCKITE SUITE

When Holland (1900) coined the term "charnockite" he was reluctant to have it applied outside India. Nevertheless, under a misconception that the Madras-type rock formed an homogeneous hypersthene-bearing pluton, hypersthene-bearing plutons described around the world have been variously termed charnockite or charnockitic. Despite the misconceptions (and the resultant confusion in the use of the term *charnockite*) such a hypersthene-bearing suite of plutonic igneous rocks does appear to constitute a homogeneous entity, having similar petrogenetic and field relationships wherever it occurs. It is most frequently found in association with the widespread middle to late Proterozoic anorthosite-rapakivi granite complexes of the continental cratons, examples being described from Quebec, Canada (Grenville and Nain provinces), New York State, U.S.A. (Adirondacks), Finland (Wiborg massif), Norway (Rogaland, Farsund, Jotunheim), Sweden (Varberg massif), U.S.S.R. (Korosten complex, Ukraine), and Australia (Arunta complex).

As a petrogenetic suite these plutonic rocks are of calc-alkaline affinities (Streckeisen, 1976) reflected in a nomenclature that uses hypersthene as a prefix to the notation of the QAP triangle (Fig. 1). However, as the various special names are already well established in the literature, these are retained as alternatives. Thus hypersthene-alkali feldspar granite is also referred to as alkali feldspar charnockite, hypersthene granite as charnockite and *farsundite*, hypersthene granodiorite as *opdalite* or *charno-enderbite*, etc. (Fig. 1). This spectrum effectively links Howie's (1955) "charnockite suite" with the anorthosite-jotunite-mangerite "Jotun Kindred" of Goldschmidt (1916, 1922). The term *monzonorite* has been used for a quartz-poor member of the suite containing more plagioclase than micropertthite and *M-enderbite* for the member containing both mesoperthite and plagioclase.

Diagnostic characteristics

All members of this suite are orthopyroxene-bearing or carry fayalite-quartz pairs; the orthopyroxene is usually hypersthene. Feldspars are alkalic, usually mesoperthite, or in the range albite-labradorite when no K-feldspar is present. When quartz is present it is invariably colored. An abundance of Ti is indicated by features such as sagenitic exsolutions in biotite and the common presence of sphene, ilmenite, and rutile as accessory minerals. The anorthosites often carry massive ilmenite or titaniferous

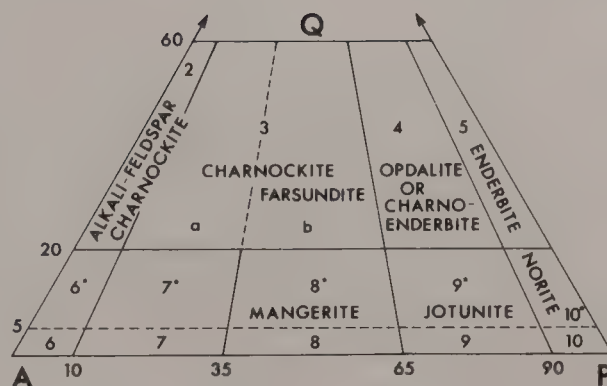


FIGURE 1. Mineralogical classification of the charnockite suite; fields in the QAP diagram are given the traditional names; the IUGS recommended names are: 2, hypersthene-alkali-feldspar granite; 3a,b, hypersthene granite; 4, hypersthene granodiorite; 5, hypersthene tonalite; 6, hypersthene-alkali-feldspar syenite (6* quartz-), 7, hypersthene syenite (7* quartz-); 8, hypersthene monzonite (8* quartz-); 9, hypersthene monzodiorite (9* quartz-); 10, hypersthene diorite or hypersthene gabbro (10* quartz-) (where $M < 10$ = anorthosite). (After Streckeisen, 1976.)

magnetite segregations of economic importance (e.g., Egersund, S Norway). Garnet, usually a pyrope-rich member, is also a widespread phase. Garnet and amphibole may form corona structures around orthopyroxene megacrysts.

The granulite paragenesis and field association has led to much controversy concerning whether these rocks are plutons emplaced and crystallized in the granulite facies or plutons cooled and crystallized prior to granulite facies metamorphism.

Petrogenesis

The origin of the parent magmas to the massive anorthosites is germane to this problem. Crustal melting and massive crustal assimilation have both been popular models in their time, but isotopic studies seem to indicate a mainly mantle source, probably an alkaline basalt melt (Morse, 1982). The development of the gabbro-anorthosite suite from this parent melt is generally held to be the result of fractional crystallization. However, the relationship of these rocks to the jotunite-charnockite suite remains more controversial. Two models are currently being evaluated, largely based on studies of the Varberg massif, Sweden (Hubbard and Whitley, 1978, 1979) and in Colorado, U.S.A. (Barker et al., 1975).

Model 1 considers the contamination of a gabbro-anorthosite magma series by an assimilated and partially melted lower crustal slab. This is considered to produce a progressively more siliceous suite of magmas subsequently

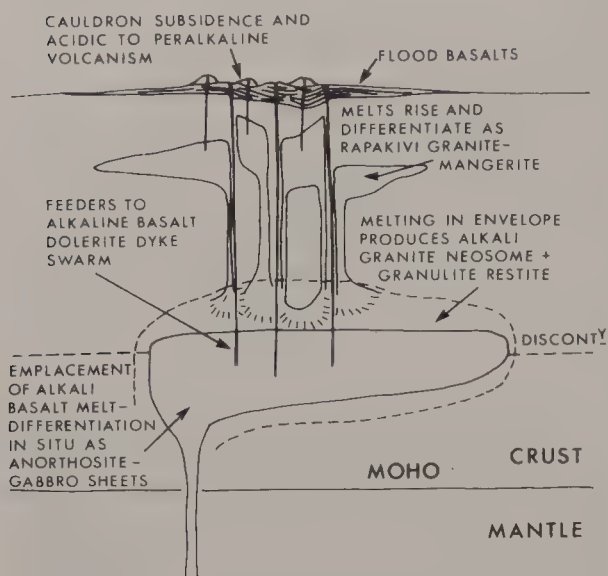


FIGURE 2. Schematic relationship between rocks of the charnockite suite, the gabbro-anorthosites and rapakivi granite-mangerite-alkaline basalt association. (Based on Barker et al., 1975; Hubbard and Whitley, 1979.)

intruded at very deep levels. Model 2 considers a partial melting of the lower crustal envelope to a gabbro-anorthosite intrusion at the base of the crust (Fig. 2). The first partial melt is virtually a eutectic neosome with a composition close to charnockite (and physically very similar to Holland's (1900) original type), while progressively greater melt fractions define a trend towards enderbite and jotunite. The more basic magmas remain to be emplaced deep in the crust, as do the volumetrically small eutectic melts, but the more buoyant and abundant intermediate farsunditic to mangeritic melts rise to higher levels giving rise to the rapakivi granite-mangerite association. This model is also consistent with the REE chemistry from the Varberg massif (see **Geochemical modeling in petrogenesis**) but, on the other hand, the Indian, Australian, and Swedish examples do not have associated anorthosite plutons.

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Cross-references: *Anorthosite*; *Charnockite*; *Geochemical modeling in petrogenesis*; *Granulite facies*; *Plutonic rocks—classification and nomenclature*; *Rapakivi texture*; *Volatiles*.

CIRCUM-PACIFIC PLUTONIC AND VOLCANIC ACTIVITY

Plutonic activity

Plutonic rocks are ubiquitous around the circum-Pacific margin, although hidden in many areas beneath volcanic and sedimentary rocks. The granitoid rocks form the base for the great circum-Pacific "ring of fire," and form most of the continental crust, which varies from <20 km to >70 km thick beneath the Andes, probably the thickest in the world (James, 1971).

Owing to its enormous complexity, circum-Pacific plutonism is not easily summarized. Unfortunately for simple concepts, the Pacific Ocean basin does not unify the diverse marginal terranes in any significant way except possibly for potassium variation perpendicular to the basin margins. Except for Australia and New Zealand, the circum-Pacific granitic rocks are commonly more basic, more heterogenous, and more foliated on the oceanic side. Towards the continents, K_2O and the $K_2O:SiO_2$ ratio increase (cf., Sierra Nevada Batholith—Bateman et al., 1963). The most abundant plutonic rock type around the Pacific Ocean is granodiorite, followed by quartz diorite and quartz monzonite or adamellite. Diorite is

TABLE 1. Main periods of plutonic activity in the circum-Pacific region

	Chile	Peru	Mexico	Sierra Nevada Bath.	Cascades	Coast Range Bath.	Alaska	Japan	N Zld	Aust.
L. Tertiary		×			×	×	×	×		
E. Tertiary	×	×	×		×	×	×	×		
L. Cretaceous	×	×	×	×	×	×	×	×		
E. Cretaceous	×			×		×		×		
L. Jurassic	×	×	×	×		×	×			
E. Jurassic							×	×		
L. Triassic						×			×	
E. Triassic										
L. Palaeozoic	×		×						×	×
E. Palaeozoic							×			×
Precambrian	×	×	×		×					

common and gabbro much less so; true granite is rare.

Although the great Mesozoic batholiths are the best known and characterize the circum-Pacific belt, plutonism spans a much longer time range. Precambrian plutonic rocks are known in SW United States, N Mexico, and Peru, and suspected in many other places. The Palaeozoic is represented in Australia by batholiths fully comparable in size to the largest of the Mesozoic masses on the eastern side of the Pacific, and by plutons in SE Alaska, S Mexico, and Chile. Tertiary plutons are second only to those of the Mesozoic in abundance. The age of emplacement of the plutonic rocks (Table 1), which in many cases should be distinguished from the time of origin, commonly decreases toward the continents within the marginal zone, especially on the eastern side of the Pacific basin, e.g., the Coast Plutonic Complex, Canada, the Sierra Nevada Batholith, U.S.A., and in the Chilean Andes—(Fig. 1). Roughly parallel with this transgression is an increase in alkalinity and silica content, conforming on a regional scale with the basic to acidic sequence of emplacement commonly seen within small areas.

Emplacement ages of most Japanese plutonic rocks are clustered in two groups of belts separated by a transverse rupture known as the Fossa Magna (Fig. 2). The four belts of plutonic rocks NE of this lineament show a progressive increase in age from Miocene and younger just E of the rupture, to Early Cretaceous at the northeastern end of Honshu and SW Hokkaido. Most of the younger intrusive rocks occur in SW Japan and, except for those in the Hida district, none is older than 100 Ma.

The most striking feature of the great Coastal Batholith of Peru is the centered complexes that are regularly spaced about 80 km apart along the axis of the batholith. The arcuate plutonic masses which form the complexes appear to represent high-level intrusions that may be subvolcanic and may have been accompanied by cauldron subsidence (Pitcher, 1978).

Recent studies, mainly in NW Mexico by Gastil (1983), indicate that a volcanic–plutonic arc extended from northern Sonora through what is now the peninsula of southern and Baja California to S Mexico. The arc is particularly interesting in that its western edge crossed oceanic sea floor and its interior edge penetrated Precambrian continental terrane.

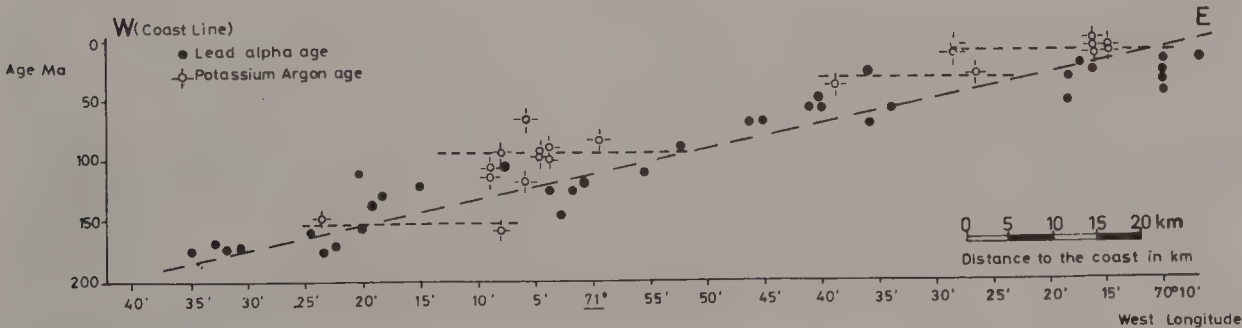


FIGURE 1. Increase in age of granitic rocks from W to E in Chile. (From Aguirre et al., 1974.)



FIGURE 2. Age grouping of plutonic rocks in Japan. (After Kawano and Ueda, 1967.)

The belt is divided, from W to E, into three segments characterized, respectively, by gabbro, tonalite, and adamellite, the position of each apparently being related to the type of pre-existing crust. Intrusion began about 145 Ma ago on the Pacific side and moved across the belt in about 60 Ma.

The Sierra Nevada Batholith forms a huge block (800 km long) which was tilted westward (opposite to the Andes) in Cenozoic time. The batholith was emplaced into strongly deformed but only weakly or moderately metamorphosed Palaeozoic and Mesozoic strata, approximately along a line separating an eastern miogeosyncline from a western eugeosyncline. The predominance of inward-facing tops and the progressively younger age of pendant strata towards the center of the batholith indicate that it occupies the site of a complexly faulted synclinorium. Granodiorite and quartz monzonite, about equally divided, make up most of the plutonic terrane (Bateman et al., 1963).

The Coast Plutonic Complex, one of the largest post-Precambrian batholithic complexes in the world, extends along the western side of British Columbia for a distance of about 1,760 km from near the Canada-United States border to the Yukon (Roddick, 1974). Essentially, it consists of more or less well-defined plutons and pendants of metavolcanic and

TABLE 2. Average chemical analyses of common plutonic rocks from the Coast Range Batholith, British Columbia, Canada

	1	2	3	4	5
SiO ₂	49.75	59.03	63.48	69.60	57.33
TiO ₂	1.06	0.70	0.53	0.28	1.04
Al ₂ O ₃	17.38	17.01	16.48	15.62	16.24
Fe ₂ O ₃	2.97	2.06	1.33	0.77	2.11
FeO	5.62	3.51	2.65	1.52	4.98
MnO	0.15	0.12	0.09	0.06	0.14
MgO	5.45	2.80	1.99	0.88	3.60
CaO	8.79	6.05	4.51	2.82	6.97
Na ₂ O	4.67	4.80	4.50	4.39	3.76
K ₂ O	1.10	1.69	2.28	3.07	2.22
P ₂ O ₅	0.33	0.25	0.20	0.12	0.28
S	0.04	0.01	0.01	0.01	0.21
H ₂ O	1.47	1.02	0.86	0.56	1.15
CO ₂	0.03	0.05	0.03	0.03	0.28

1, 13 diorites; 2, 68 quartz diorites; 3, 63 granodiorites; 4, 12 quartz monzonites; 5, 25 granitoid gneisses.

metasedimentary rocks, all embedded in a heterogeneous matrix of plutonic rocks, granitoid gneisses, and migmatites. About half of the batholithic complex is quartz diorite and granodiorite, roughly equally divided. Diorite and dioritic migmatites form about 15% of the complex and are concentrated in the western part of the belt. The Coast Plutonic Complex is the most basic and complex of all major plutonic terranes along the eastern margin of the Pacific Ocean basin. Average chemical composition of the common rock types is shown in Table 2 and representative modes in Fig. 3. The latter shows the strong concentration in the quartz diorite-diorite fields that is a dominant characteristic of the belt.

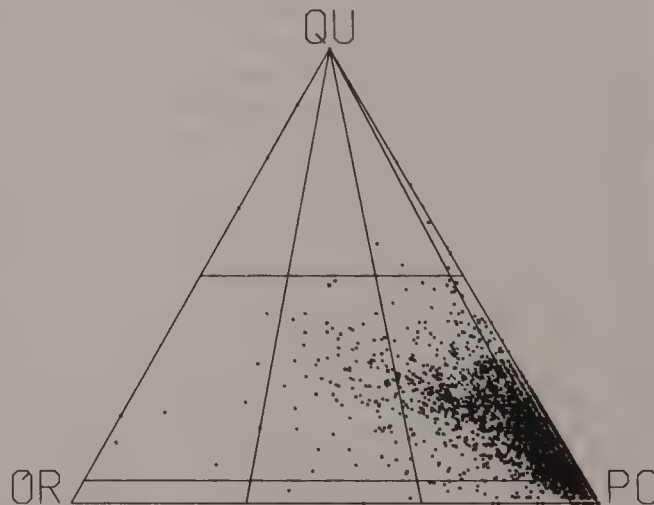


FIGURE 3. Plot of modes (estimated from sawn surfaces) of 3,155 specimens from Coast Range Batholith, Canada.

Volcanic activity

The great circum-Pacific "ring of fire" has two sides, eastern and western, and they differ. A system of island arcs borders the whole eastern side of Asia, then swings eastward before extending southward to New Zealand (Fig. 4). Along the Pacific Coast there are no true island arcs but instead a series of continental-margin volcanic chains. The Aleutian arc connects the volcanic zones of Asia and the Americas. Major orogenic systems on the eastern side of the Pacific Ocean are parallel to the basin, but the western side is complicated by the intersection of Eurasian systems with circum-Pacific belts, a



FIGURE 4. Trenches and ridges in the western Pacific Ocean: 1, Aleutian; 2, Kurile-Kamchatka; 3, Japan; 4, Nansei; 5, Philippine; 6, Izu-Bonin; 7, Marianas; 8, Yap; 9, Palau; 10, New Guinea; 11, West Melanesian; 12, New Britain; 13, Solomon Islands; 14, East Melanesian; 15, New Hebrides; 16, New Caledonia; 17, Tonga; 18, Kermadec; 19, New Zealand; 20, Hawaiian; 21, Marcus-Wake; 22, Eauripik; 23, Caroline; 24, Kapingamirangi; 25, Marshall Islands; 26, Tokelau and Cook Islands; 27, Line Islands; 28, Northwestern Pacific mountains; 29, Bering Sea; 30, Sea of Okhotsk; 31, Sea of Japan; 32, East China Sea; 33, South China Sea; 34, Sulu Sea; 35, Celebes Sea; 36, Molucca Sea; 37, Banda Sea; 38, Bismarck Sea; 39, Solomon Sea; 40, Coral Sea; 41, northern basin of Fiji Sea; 42, southern basin of Fiji Sea; 43, New Caledonian basin; 44, Tasman Sea; 45, Philippine oceanic basin. (After Garshkov, 1970.)

fact expressed by the complex arrangement of land masses.

In most island arcs the volcanic rocks show an increase in alkalis away from the ocean basin. The normal basalts of island arcs, with $\text{Na}_2\text{O} + \text{K}_2\text{O} = 2-3\%$, are replaced on the continental platform (or deeper within it) by alkalic basalts and trachybasalts with $\text{Na}_2\text{O} + \text{K}_2\text{O} = 6-8\%$; there is a corresponding replacement of the acidic dacites and rhyolites ($\text{Na}_2\text{O} + \text{K}_2\text{O} = 6-8\%$) by trachytes and phonolites ($\text{Na}_2\text{O} + \text{K}_2\text{O} = 12-14\%$). The change in composition of the lavas from calc-alkalic to alkalic also occurs during the development from geosyncline to platform. For example, geosynclinal lavas of the Mesozoic in NE Asia are calc-alkaline but modern lavas there are purely alkalic.

The most persistent trenches and deep earthquake belts generally coincide with the "andesite line." One of the principal problems concerning circum-Pacific volcanism is the source of the most common rock type lying on the continental side of this line, namely, andesite. Geophysical evidence and geochemical inference imply, in places, that andesites have come directly from the mantle (Dickinson, 1968). The composition of andesites varies across island arcs and marginal continental volcanic chains, and if the mantle-source theory is assumed, this may be accounted for by postulating either a variable mantle composition, or differing depths of partial fusion (Kuno, 1966). The main problem encountered in deriving major amounts of andesite from a basaltic rock or magma by partial fusion or differentiation is that the proportion of basalt to andesite in andesitic chains is very small, whereas the proportion on average should be reversed. The geophysical evidence is negative, that is, it does not indicate a deep crust or appreciable sialic material beneath some island arcs that might be a source of andesite magma, or at least of contaminating sial. Taylor and White (1966) showed that the trace elements in andesites are in accord with a mantle source, but they also may be compatible with other sources.

The world's thickest volcanic piles were extruded in Jurassic and Cretaceous time in the Andes, where they probably exceed 20 km of andesitic flows and pyroclastic rocks.

The volcanoes of Mexico are unique among volcanic chains in their lack of parallelism with the abyssal trench (Central American Trench). Instead, they form a transverse trend from Ceboruco volcano on the Pacific coast to San Martin volcano (Table 3) on the Gulf of Mexico. The chain extends westward far beyond the trench, forming the volcanic islands

TABLE 3. Average lava compositions for the volcanoes of the United States of America and Mexico

	1	2	3	4	5	6	7	8	9	10	11	12
SiO ₂	53.7	61.4	68.6	74.0	51.8	61.5	69.2	74.9	48.7	55.6	60.3	48.3
TiO ₂	0.9	0.8	0.3	0.2	0.6	0.5	0.2	0.2	0.7	1.0	0.8	2.0
Al ₂ O ₃	18.2	17.2	16.2	14.2	18.4	17.1	16.2	14.3	12.9	18.3	17.4	13.3
Fe ₂ O ₃	3.8	2.4	1.2	0.7	4.2	2.8	1.8	1.0	4.9	1.6	1.4	5.1
FeO	4.6	3.3	2.1	1.3	4.9	2.8	1.2	0.6	4.5	5.8	4.4	6.6
MnO	0.1	0.1	0.1	—	0.2	0.1	0.1	0.1	0.1	0.1	0.1	0.2
MgO	5.6	3.1	1.3	0.4	6.4	3.2	1.3	0.3	9.7	5.5	3.7	9.4
CaO	8.9	6.1	3.3	1.3	10.0	6.5	3.3	1.1	12.0	7.2	6.2	9.9
Na ₂ O	3.3	4.0	4.0	3.7	2.8	3.8	4.1	3.4	2.7	3.9	4.0	3.4
K ₂ O	0.9	1.6	2.9	4.2	0.7	1.7	2.6	4.3	3.8	1.0	1.7	1.8

United States of America: *Cascade Mountains*—1, andesite-basalt (13 analyses); 2, andesite (29); 3, dacite (14); 4, rhyolite (14). *Lassen Peak*—5, basalt (5); 6, andesite (14); 7, dacite (8); 8, rhyolite (6). *Highwood Mountains*—9, phonolite (2).
Mexico: *Paricutin volcano*—10, andesite-basalt (5, 1943–44 lava); 11, andesite (1, 1952 lava). *San Martin volcano*—12, basalt (1). (After Gorshkov, 1970).

of Revillagigedo and a number of submarine volcanoes. Farther west the trend is represented by the Clarion Fracture zone. This also is a peculiarity of the Mexican chain, for none of the other E-trending fracture zones in the eastern Pacific have volcanic extensions on the continent.

The western parts of both the United States and Canada exhibit a more passive boundary between the ocean basin and the continent with transitions to active boundaries in Alaska and Central America. The mainly andesitic and basaltic volcanism of the Mesozoic in California is complex (see Table 3 for average chemical analyses). Oregon and Washington, however, are dominated by volcanic rocks and landforms of Tertiary and Quaternary age. No equivalent to this great volcanic panorama exists between Mexico and Alaska. The prolonged volcanic regime has produced a variety of lava plateaus, volcanic cones, and eruptive ranges.

Analyses of Quaternary volcanic rocks show a clear increase in alkalis from the western Lassen Peak calc-alkaline rocks to the Highwood Mountains alkaline rocks (Table 3).

The volcanoes of the Aleutian arc are unusual in that they reveal a clear longitudinal zonation in petrochemistry. At the western end (Rat Islands) the basalts and andesites are similar to the Lassen Peak type, but the dacites are enriched in alkalis whereas the rhyolites contain less alkalis. At the eastern end, the variation curves fall between the Lassen Peak and Yellowstone types.

Generally, the volcanic belts that extend along the margin of Asia mark the boundaries of Cenozoic geosynclinal zones and are linked only in a casual way to the Pacific Ocean basin. Extremely diverse compositions are obscured

within different belts, the product of a number of volcanic cycles. The belts are not unique to the circum-Pacific area, but are found far to the west.

J. A. RODDICK

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Cross-references: *Andesite and dacite*; *Andesite line*; *Batholith*; *Calc-alkaline rocks*; *Granodiorite*; *Igneous rocks—classification and nomenclature*; *Igneous rocks—tectonic setting*; *Island arcs—petrology*; *Plutonic rocks—classification and nomenclature*.

COLUMNAR JOINTING

Columnar joints break rock masses into long prismatic units in which the fractures along the prisms or columns are continuous and dominant, whereas those across the prisms are weaker and generally terminate at the edges of the column. The structure is most prominent in lava flows, particularly basalts (Fig. 1), but is also found in andesites, rhyolites, obsidians, and other rocks. It is less common in sills and dykes. Examples occur in sediments such as muds, gypsum, and ignimbritic tuffs. Hornfelses are rarely columnar except for the high-temperature buchite.

In cross-section most columns show hexagonal symmetry and some rocks are broken into neatly fitting 6-sided prisms with equal sides and equal angles. Such perfection, however, is by no means the rule; 5- and 7-sided columns are common and 4- and 3-sided examples are not rare.

Columns in basalt are most commonly about 60 cm across; they reach about 2 m in some thick dolerite sills but columns 4-5 m across have been found in thin basalt flows by the Columbia River, Oregon, U.S.A. Slender columns a few centimeters across or less seem to be more typical of tuffs and buchites.

The most perfect columns have smooth and parallel sides and straight axes, but others have curved axes; changes in prism diameter (*pinch and swell structure*) are common. Grooves on the joints perpendicular to the axis are common and are called *chisel marks* (James, 1920). The prisms are broken by cross-fractures, which may

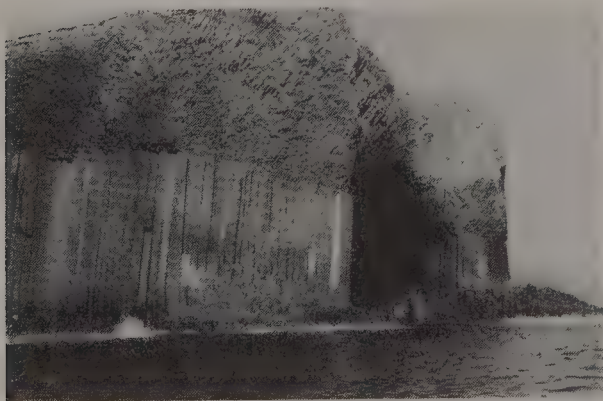


FIGURE 1. Columnar jointing in basalt, Fingal's Cave, NW Scotland.

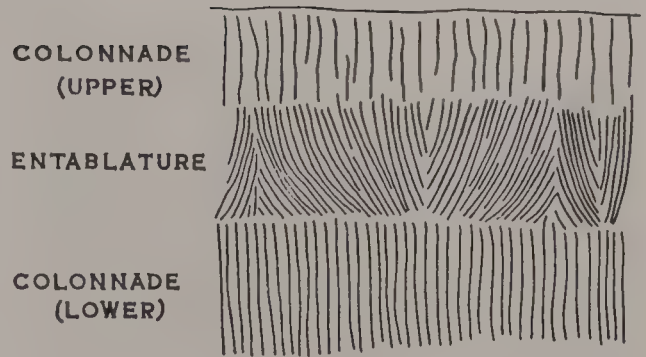


FIGURE 2. Three-fold division of a basalt flow with the upper zone regularly columnar; Columbia River, Oregon, U.S.A. (From Spry, 1962.)

be planar or curved to give *ball and socket joints* (*cup and ball structure*). Weathering of columnar basalt with closely-spaced cross-fractures gives a "pile of Dutch cheeses."

A three-fold division in the joint pattern is common (Tomkiew, 1940; Spry, 1962). The lower part (Fig. 2) contains regular vertical columns and is called the *colonnade*. The central zone is characterized by irregular twisted columns and is named the *entablature*. The upper zone may be regularly columnar (*upper colonnade*), crudely columnar (*pseudo-columnar*), or structureless.

The curved and twisted columns of the entablature contain a number of common patterns, e.g., chevron (upright, oblique or inverted), rosette, fan (upright or inverted), and basin. In some flows the entablature is closely and irregularly fractured with *brickbat jointing* and has been called *diced lava*.

The joint patterns in many bodies are produced over a period of time, with master joints first, then perhaps megacolumns, then the columns themselves (colonnades first and entablature last) and finally the cross-joints. Part of this sequence has been observed in lava lakes at Kilauea (Peck and Minakami, 1968).

Columnar jointing in igneous rocks has been attributed to convection and concretion during crystallization, but the accepted hypothesis invokes tensional rupture due to the thermal stresses following contraction of the solidified igneous rock. The liquid magma contracts as it cools but the viscosity rises at lower temperatures and contraction takes place first by liquid flow, then by creep; as the magma solidifies the rate of contraction exceeds the rate of relief by creep. Thermal stresses accumulate until they exceed the tensile strength of the rock and then fracture takes place. Columnar fracture in hornfelses and tuffs has a fundamentally similar origin and that in muds results from contraction following dehydration. Lachenbuch (1962) has

shown that the polygonal fractures in permafrost are due to thermal contraction and ice-wedging; much of his discussion is applicable to all columnar jointing.

Jointing depends on thermal stresses that arise because free change of volume of each volume-unit cannot take place. If the temperature of a body is changed, no stress occurs if it is homogeneous, isotropic, and unconstrained. Stresses develop if the body is prevented from expanding or contracting, or if there is an uneven temperature distribution. The magnitude of the thermal stresses in a cooling igneous body depends on physical properties such as the coefficient of expansion, Young's modulus, Poisson's ratio, and the conductivity, and on the presence of existing fractures.

Fractures are initiated on the outer surfaces of the body and extend inwards as the temperature decreases; thus, the hexagonal pattern is first a surface fracture pattern. Two hypotheses have been advanced. One explanation involves production of the pattern by the essentially instantaneous development of fractures around equally spaced "centers of cooling", the hexagonal form being the result of the principle of least work. Another explanation (Spry, 1962) envisages the continuous propagation of fractures that bifurcate as they grow, similar to fractures in glass.

Once the fracture pattern is formed at the surface, further cooling will produce additional thermal stress and the fractures penetrate inwards perpendicular to the surfaces of equal thermal stress. In bodies of simple shape, the isotherms are parallel to the surfaces of equal stress and thus the columns are perpendicular to the isotherms and to the external boundary of the body (Jaeger, 1961). This is not necessarily so, particularly in the entablature or near marked irregularities in the boundary, or in the vicinity of master joints.

Two distinct patterns of columns appear to occur in volcanic necks. At high levels (e.g., Devil's Tower, Wyoming, U.S.A.) an inverted fan is formed, but deep within the conduit a rosette with a vertical axis is formed (e.g., Rock and Spindle, St Andrews, Scotland).

Columnar jointing in hornfelses (Spry and Solomon, 1964) is due to thermal stress produced by contraction during cooling. Contraction may be due to dehydration of hydrous phases, recrystallization and destruction of pore space, or the production of new minerals of higher density. Columnar structure is marked in buchites where the production of a liquid phase (due to partial melting) allows considerable volume change by repacking of grains and filling of pore-space.

ALAN SPRY

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Cross-references: *Buchite; Flood basalt; Lava and lava eruptions; Volcanic neck.*

CONE SHEETS

Cone sheets are minor intrusions arranged in a curving array or set, individually inclined rather steeply towards a common focal point (Fig. 1). The upward and outward splay of the sheets entails an annular plan for a set as a unit, or an arcuate plan where the sheets are partially developed; there is a central area, over the focus, devoid of sheets. Cone sheets are a feature developed in most of the intrusive centers of the British Tertiary Volcanic Province (Richey, 1932; Walker, 1975) where the thickness of individual cone sheets remains generally constant throughout a particular set but may vary within wide limits in different sets. For example, on Mull and Ardnamurchan in the Hebridean area, some sheets around the earliest centers are 10-12 m thick, whereas in certain later sets thicknesses of 2-6 m are usual.

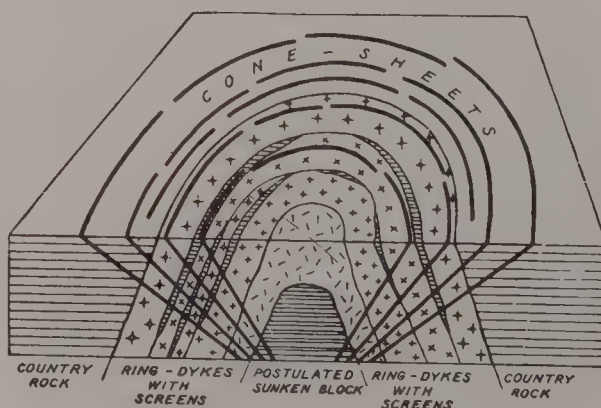


FIGURE 1. Ideal ring complex of ring dykes and cone sheets. (From Richey, 1932.)

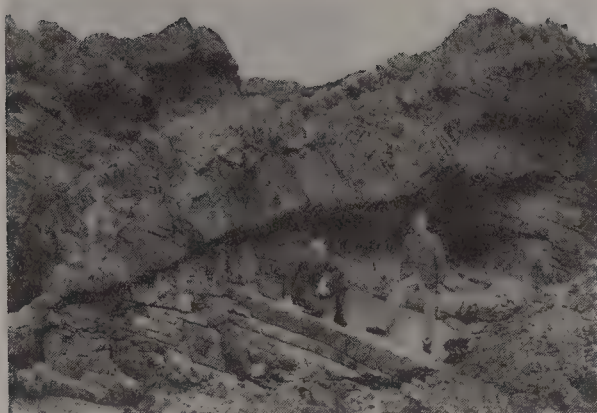


FIGURE 2. Cone sheets of the outer set of Centre 2, Ardnamurchan, W Scotland, at Kilchoan; country rock is siliceous schist.

Rock types. Cone sheets are exposed in denuded volcanic centers, and as such the rock types that form them span most of the range of volcanic and hypabyssal compositions. A feature of cone sheets in the British Tertiary Volcanic Province is that many are composite,

with basic margins and intermediate or acidic centers.

Inclination and depth of magma reservoir. The average inclination of sheets varies with different sets, but the commonest angles range from 30° to 45° (Fig. 2). There is sometimes a tendency for the inclination to be lower towards the outer side of the set and higher towards the center. This is most striking in Ardnamurchan, where the outermost sheets around Centre 1 (Fig. 3) are inclined at only 10–15° and those of the inner set of Centre 2 dip at 65–70°, a much greater angle of inclination than is found in any other district (Richey et al., 1930).

In the Kila–Warji complex, Nigeria, dips of 30–45° would bring cone sheets to a common focus 5.6–6.4 km below the present surface (Turner, 1968). In the British Tertiary Volcanic Province the depth of focus would be nearer 4.5–5 km, but account has to be taken of the amount of crustal material that has been eroded from above the intrusive complex. Taking this into account, a magma chamber 6.5–7 km down at the time of emplacement is postulated.

Numbers. The number of sheets composing a set varies from place to place around a

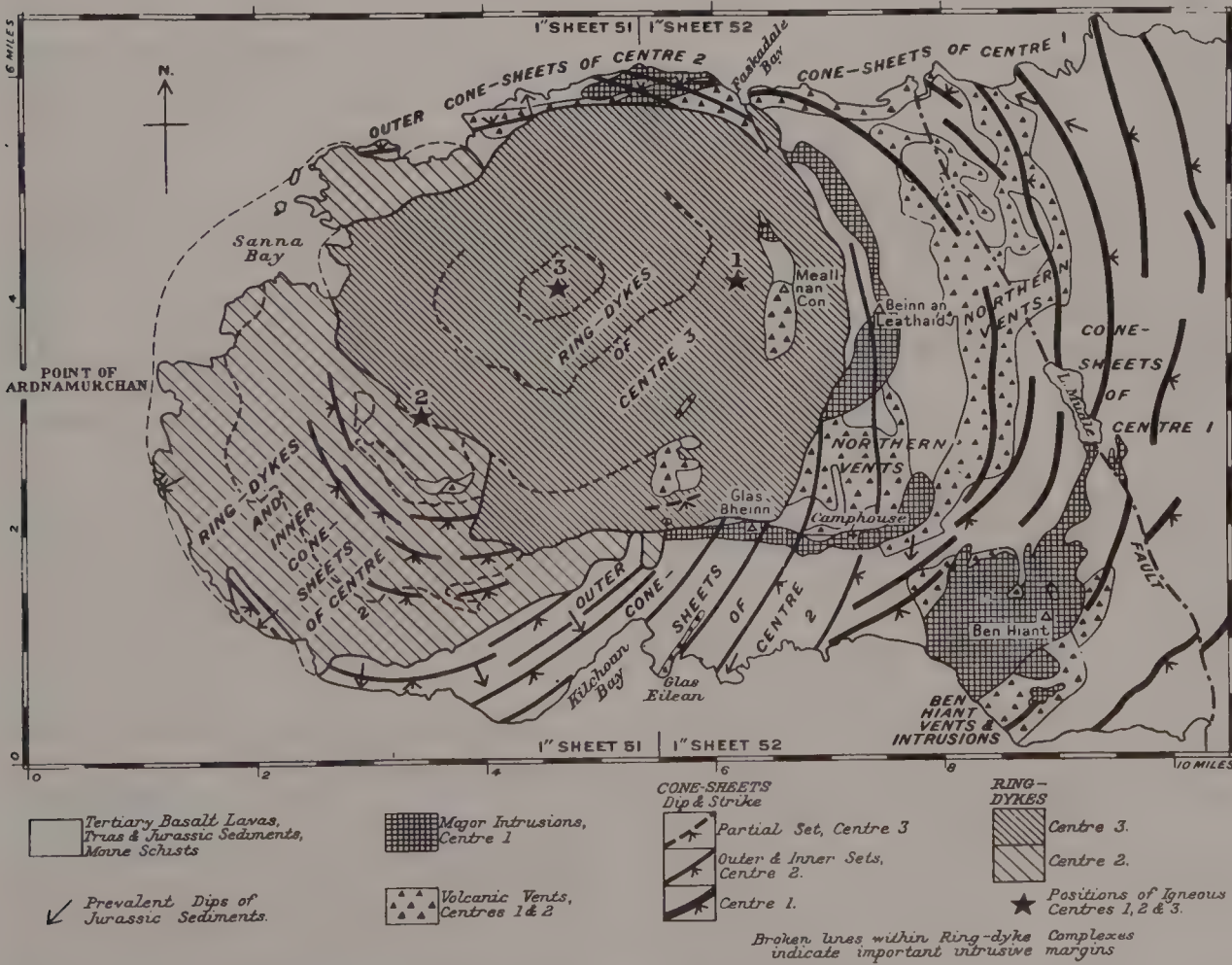


FIGURE 3. Ring complexes of Ardnamurchan, W Scotland. (From Richey et al., 1930.)

province, as well as in different sets. An extreme example is the concentrated intrusion of sheets that occurs on Mull, where almost an entire mountain (Ben Talaidh), over 800 m high, is composed of cone sheets with only occasional narrow strips of the earlier rocks remaining. On the other hand, at Sleive Gullion in Ireland, the total number of cone sheets is no more than half a dozen.

It is considered that space for the sheets was provided by raising the roof of each individual intrusion when it was emplaced (cf., Walker, 1975). In the case of the late basic set of cone sheets on Mull, this would have involved a central uplift of some 900 m. An even greater central uplift has been calculated for the outer set of Centre 2, Ardnamurchan.

Manner of emplacement. The factors that might give an indication of the way in which cone sheets are emplaced include the depth of the magma reservoir feeding the intrusions, the nature of the magma, and the stress system operating in the crust at the time of intrusion. The depth of the top of the magma source is estimated generally at deeper than 6.5 km (at the time of intrusion), an exception being the unusual cone sheet-like structures associated with the Homa Mountain carbonatite complex in Kenya (King et al., 1972), where the focus lies at a depth of 1–2 km, and the cone sheets appear to have been emplaced by explosive activity. In all other cases the focus is much deeper and the emplacement has to be related to more widespread crustal stresses.

The nature of the magma will determine the density contrast between it and the country rock, and hence the excess of magma pressure over vertical stress in the crust. In the case of a basaltic magma with an assumed density of 2.9

and a magma source at a depth of 58.2 km within the upper mantle (Roberts, 1970), the excess pressure at a depth of about 7 km would be approximately 1 kb. The pattern of stress surrounding a magma chamber in these circumstances is difficult to model accurately, but simplified models (Fig. 4) appear to indicate the likely occurrence of a pattern of tensile fractures which, if filled by magma, would result in the observed characteristics of cone sheets (Phillips, 1974). The formation of tensile fractures would depend on the tensile failure of the crust and where this did not take place then there might simply be a bodily elevation of the central block within a ring complex (cf., Walker, 1975). That this seems to happen may be confirmed by the evidence of resurgence seen in some ring complexes and surface calderas.

J. G. MACDONALD

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Cross-references: *British Tertiary Volcanic Province*; *Calderas and volcano-tectonic depressions*; *Dyke (dike)*; *Emplacement-modes*; *Hypabyssal rocks*; *Ring dyke*.

CONTACT (THERMAL) METAMORPHISM

Contact (thermal) metamorphism is the phenomenon of recrystallization and re-equilibration seen in the country rocks adjacent to intrusive igneous bodies. Contact metamorphic effects can be associated with intrusive rocks of any composition, and in any emplacement mode

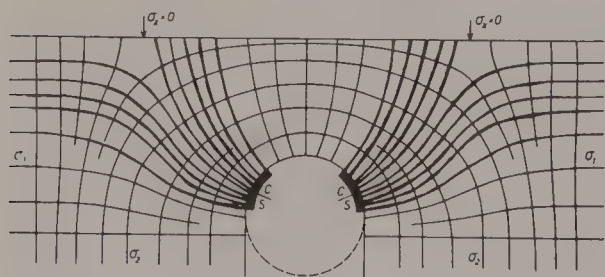


FIGURE 4. Principal stress trajectories around a circular cavity near the free edge of a semi-infinite plate. The sectors marked C and S represent the sectors where the tangential stress at the surface of the cavity has a minimum value under conditions of excess internal pressure; the heavy lines represent sheet intrusions that would form parallel to the directions of maximum principal stress under such conditions, cone sheets developing from tensile failure in sector C, and sills resulting from tensile failure in sector S. (From Roberts, 1970.)

(e.g., adjacent to dykes, sills, or major batholiths). Similar effects may also occur beneath thick lava flows. The nature of contact metamorphism varies from textural re-equilibration to major petrographic change that may involve bulk chemical alteration, partial or total melting (metasomatism, anatexis). The scale of the phenomenon can range from a zone only millimeters to several kilometers wide. Such contact metamorphic zones are termed *aureoles*.

Controls

When a pluton intrudes cooler country rock, the degree of change reflects the rate of cooling in the pluton (a function of heat capacity), the heat capacity of the country rock, and the heat flow in the country rock (itself a function of cooling mechanism, heat capacity, and conductivity). When intrusion is followed by cooling in situ and solidification of the magma, the arrival of the heat source can be treated as being instantaneous; or, in a more complex situation, when the magmas arrive in pulses, as a series of instantaneous events. Cooling itself occurs by conductive heat flow, fluid convection, or a combination of the two processes. The result is a thermal gradient whose shape reflects the particular process involved. In an instantaneous situation, the pluton is a decaying heat source, but if the pluton represents a conduit through which magma rose to higher levels it may have provided a nondecaying heat source for a considerable length of time. A pluton that cooled primarily by conductive heat loss will be marked by a narrow, but intense thermal aureole, possibly including an internal zone marked by partial or total fusion (buchite development), ultrametamorphism, or pyrometamorphism. Such phenomena suggest that the highest temperatures reached in the country rock adjacent to the pluton may have approached close to those of the magma liquidus.

Around a pluton where convective cooling was the dominant mechanism such high temperatures may never have been approached, but depending on the porosity profile and permeability of the country rocks the thermal aureole may be very broad. In cases of very high porosity and permeability, convective cooling may have proved so efficient that thermal alteration is minimal. Instead, the circulation of the vast amounts of fluid involved in such systems can result in wholesale metasomatic changes (cf., Einsele et al., 1980). Such aureoles may contain extensive mineralization (e.g., porphyry-Cu and -Mo deposits—see **Granites and mineralization**).

Influence of emplacement mechanism

The size of an intrusion has an obvious effect on contact metamorphic phenomena; typically, small hypabyssal intrusions are associated with little, if any, contact metamorphic effects. Thus, small sills, dykes, cone sheets, and such bodies have restricted thermal aureoles. Feeder dykes and volcanic necks and plugs have aureoles of highly varied size, and quite diverse contact metamorphic effects are associated with them. The former may contain engulfed xenoliths that may display the effects of pyrometasomatic change—buchite, or sanidinite facies mineral assemblages.

With large intrusions, emplacement mechanisms can influence the form of the contact metamorphic effects considerably. Permissive (permitted, passive) intrusion by brittle fracturing and piecemeal or large-scale stoping, contrasted with diapiric, forceful (forcible) intrusion associated with ductile deformation, represent two ends of a spectrum. In the case of the former, high-level emplacement may involve cauldron-subsidence, with large amounts of heat released during volcanic outbursts and the escape of dissolved volatiles. In such cases, large plutons may have small aureoles and negligible contact metamorphic effects, despite there being a steep thermal gradient between pluton and country rock.

Around minor intrusions and permissively emplaced plutons, textural reequilibration in the country rock usually involves static mineral growth resulting in an isotropic grain fabric—often with hornfelsic texture. Where forceful intrusion is involved, ductile deformation can involve the imposition of an anisotropic grain fabric or schistosity. The products of this form of contact metamorphism may resemble those of regional metamorphism, differing only in scale of development. The interaction of such contrasting emplacement mechanisms with forms of contact metamorphism is well displayed around the various granite plutons of Donegal, Eire (Pitcher and Berger, 1972—see **Aureole**).

Mineralogical and textural changes

The thermal reequilibration in the country rock consequent upon intrusion manifests itself in textural and mineralogical changes. The nature of these changes is governed by the lithostatic pressure (dependent on the crustal depth), temperature, and the activity of any mobile components such as water, halogens, CO₂, and hydrothermal electrolytes. At their simplest, such changes are reflected by a reordering of grain boundaries and grain growth. This is usually isotropic, producing

granoblastic textures typical of hornfels. Pre-existing features like bedding, fossils and cleavage may be destroyed by a textural homogenization unless these features already constitute compositional inhomogeneities. In this latter case, in the absence of bulk movement of constituents, these features are "baked-into" the rock as palimpsest structures or textures. Where forceful intrusion has occurred, a new anisotropy may be imposed, giving the rock a schistose appearance.

Except in the case of forceful intrusion, contact metamorphism may be regarded as isobaric across a thermal gradient, the existence of which is reflected by a regular zonation with the progressive development of distinct mineral facies related to proximity to the heat source. The development of these facies depends on the composition of the country rock, and as in the case of regional metamorphism it is pelites, argillaceous limestones, and basic igneous rocks that contain the most distinctive mineral assemblages (see **Facies of thermal metamorphism**).

Where fluid movement produced bulk migration of material, metasomatism accompanies contact metamorphism. This may be confined to the country rock, or may have involved the exchange of material between the country rock and the pluton. Where such processes affect limestone the resultant rock is termed *skarn*, or, when Fe-rich, *tactite* (U.S.A.). B- and F-rich high-temperature hydrothermal fluids may cause tourmaline growth (*tourmalinization*), axinite growth, or the replacement of alkali feldspar by white mica (*greisenization*). Alkali enrichment in the country rock also goes by a variety of names, usually reflecting the action of hot brines, like the growth of white mica (propylitic alteration—*propylitization*), the replacement of Ca-plagioclase and alkali feldspar by albite (*albitization*), and Na-metasomatism (*fenitization*).

Other terms

Caustic (optalic) metamorphism—indurating, baking, burning and fritting of rocks adjacent to flows and small dykes.

Exogene effect—the effect of an igneous mass on the rock it invades.

Exomorphism—modification of country rock by invading magma or lava, i.e., contact metamorphism.

Paroptesis—a little used term for changes produced by dry heat.

Pyrometamorphism—an intense type of thermal metamorphism at temperatures near the melting points of the component minerals (see **Pyrometamorphism**).

Pyromorphism—an obsolete synonym of thermal metamorphism.

Thermometamorphism—thermal metamorphism.

ADRIAN F. PARK

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Cross-references: *Alteration, replacement; Aureole; Batholith; Buchite; Contact metasomatism; Facies of thermal metamorphism; Fenite; Granites and mineralization; Greisen; Hornfels; Hornfelsic textures; Igneous intrusions—structural behavior; Metasomatism; Pneumatolysis; Pyrometamorphism; Pyrometasomatism; Regional metamorphism; Skarn.*

CONTACT METASOMATISM

The term *contact metasomatism*, generally attributed in the first instance to Barrell (1907), refers to a process of chemical change in the composition of rocks adjacent to igneous intrusions, the change being brought about by migration of elements originating from the magma or the host. Migration may take place either by solid diffusion or through the agency of pore fluids, vapors, or gases. The term pneumatolytic metamorphism is used for contact metamorphism accompanied by strong metasomatism resulting from the chemical action of magmatic gases on both the intrusive and host rock.

Solid diffusion involves movement of ions through crystal lattices (volume diffusion) or along grain boundaries or other surfaces such as cracks across a lattice or crystal aggregate. Diffusion is controlled by crystal structure, atomic sizes, lattice defects, and surface energies. It is therefore enhanced by high temperature, which activates defects, dislocations, and stored strain energy to bring about an increase in the kinetic energy of the system. Confining pressure and deviatoric stresses, the latter being of variable development in aureoles, may also be significant in the activation of diffusion phenomena. Metallurgical studies suggest that ion migration along

grain boundaries or surfaces may be an important diffusion process, but in aureoles it is probably significant only on the scale of individual crystals, for the simple reason that the kinetic barrier to the movement of an ion along a grain boundary is greater than that to movement through a liquid or gas. Ionic migration through the agency of pore fluids is therefore likely to be the most important process in contact metasomatism. The effectiveness of ion transport in fluids is a function of temperature and pressure gradients, volume of fluid, and its chemical potential and rock permeability.

Examples of effects commonly attributed to contact metasomatism include the formation of skarns, whereby calcareous host rocks may be transformed over large areas to assemblages of coarse-grained calc-silicate minerals such as grossularite, diopside, and scapolite. Such transformation may, however, involve only very local chemical redistributions and need not necessitate exchange of a great deal of introduced elements. Formation of grossular, prehnite, and idocrase in noncalcareous rocks adjacent to intrusions of ultrabasic rocks are among other well-known effects of calcium metasomatism. Noncalcareous, pelitic or semi-pelitic host rocks in general show relatively few effects of contact metasomatism, although the introduction of boron, giving rise to tourmaline and axinite, is notable in some aureoles, for example, those surrounding the Hercynian granites of SW England. This area of Britain is also the type locality for albitization of slates, producing *adinole*s that are regarded as derivatives of alkali metasomatism from intrusive spilitic rocks; when they are banded, the term *desmotite* has been used and a *spilosite* represents an early stage of adinole formation. A metasomatic contact rock formed between granite and calcareous rocks and composed of >50% axinite, together with diopside, actinolite, zoisite, albite, and quartz is termed *limurate*. Other metasomatic effects include *fenitization* around carbonatite-alkaline intrusions, *propylitization* (propylitic alteration), which is a process of growth of white mica reflecting the action of hot brines, *tourmalinization*, *albitization*, and *greisenization* (see **Alteration, replacement**).

Metasomatic processes are locally important in economic mineralization, where hydrothermal solutions of sulfides may be introduced from an intrusion into a host such as limestone that readily undergoes solution and replacement by precipitated ores.

BRIAN CHADWICK

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Cross-references: *Alteration, replacement; Aureole; Contact (thermal) metamorphism; Fenite; Metasomatism; Pyrometasomatism; Skarn.*

CORE

The core of the Earth is its central, high-density region. With a radius of about 3,485 km, the core is thought to consist chiefly of metallic iron. The only properties of the core that can be measured directly are physical; estimates of its chemical composition are based on analogy and the need to conform with the physical parameters. At about 2,885 km depth, the velocities of P (or compressional) waves of earthquakes decrease abruptly from nearly 14 km/sec to about 8 km/sec, suggesting a change in composition from a silicate mantle of low mean atomic weight to a core of high mean atomic weight. Because S waves (transverse or shear waves) are not transmitted through the outer core, this layer is assumed to be liquid, although the innermost zone below about 5,145 km (the "inner core") is solid. For the Earth, the change in density with depth can be estimated from its mass and overall density, its moment of inertia, the variation in seismic P and S wave velocities with depth, and its free-oscillation periods (e.g., Hart et al., 1977; and reviews by Jacobs, 1975; Brown and Mussett, 1981; Bott, 1982). At about 2,885 km depth, the density is thought to increase from about 5.5 g/cm³ in the mantle to 10 g/cm³ in the outer core, reaching about 12 g/cm³ at the inner core boundary and about 12.5 g/cm³ at the Earth's center (Fig. 1).

Meteorites, which are probably representative samples of asteroid interiors, contain a metallic phase of iron with about 6% of Ni, often associated with a few percent of iron sulfide. The Earth's core is thought also to consist largely of metallic Fe with a few percent of Ni (sometimes called the "Nife" core), because of its high density and the analogy with the metal phase of meteorites. Alternative theories, that the core consists of silicate (Ramsay) or hydrogen (Kuhn and Rittmann) so highly compressed that their electronic structures had changed to metallic forms, are not supported by experiments at very high pressures and are unable to explain the density-size relationships of the inner planets (see review by Brush,

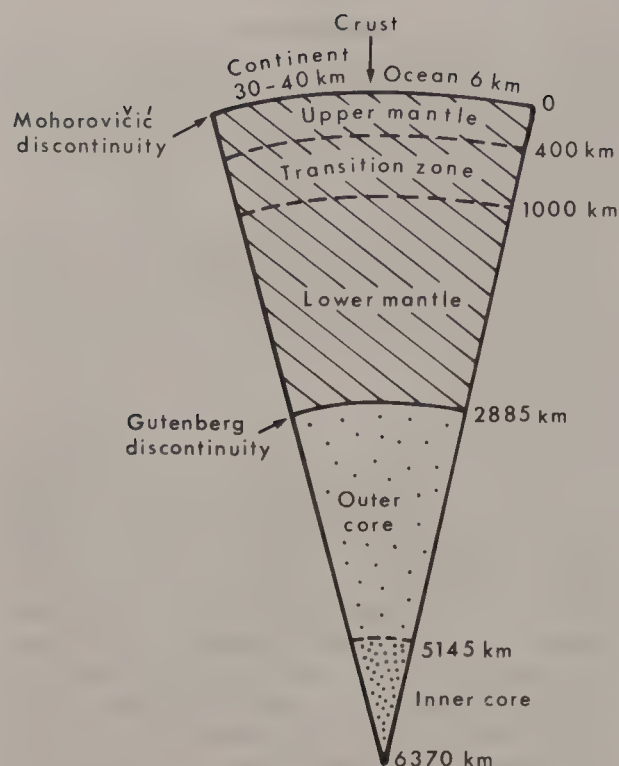


FIGURE 1. Layering within the Earth. (After Bott, 1982.)

1982). From thermodynamic considerations, and by analogy with the compositions of metal and silicate phases in meteorites and the products of metallurgical smelting processes, the metallic core should be enriched in those metals, such as Ni, Co, Cu, Ag, Au, and the Pt-group elements, that are more readily reduced than Fe to the metallic state, and in nonmetals such as P and Ge. Readily oxidized metals such as Na, Mg, Ca, Ba, Al, U, Th, Cr, etc., should be almost completely absent from the core, and confined to the silicate layers of the Earth.

The density of the core is somewhat lower than would be predicted for pure Fe or an Fe-Ni alloy at the appropriate pressures. This suggests that the core contains a proportion of elements of lower atomic number. This less-dense component has been postulated to be S as sulfide, O perhaps as FeO, or Si, dissolved in the liquid metal of the outer core (Ringwood, 1977).

By analogy with the probable melting temperature of Fe at 3.3 GPa (the pressure at the boundary between the liquid outer core and the solid inner core) the temperature of the core at this boundary is usually estimated to be between 4,000 and 5,000 K (see review by Bott, 1982). However there is some uncertainty if one considers the solid inner core to be enriched in the more refractory metals, and the liquid outer core in the more fusible components.

Fluid motions within the outer core are thought to be the cause of the Earth's magnetic field (the dynamo theory). Such motions could be due to differential rotation of the inner core and the mantle, or to thermal convection within the core. However, the latter possibility is less likely if the core is such an efficient thermal conductor that the thermal gradient within it is lower than the adiabatic gradient (see Kennedy and Higgins, 1973). If thermal convection does occur in the core, the heat source may be radiogenic, or the heat of crystallization of the solid inner core as it grows at the expense of the liquid outer core in a cooling Earth. Although a radiogenic heat source in the core is unlikely, it has been suggested that K, Rb, and Ca could be concentrated in the core as sulfides, with the radioactive isotopes ^{40}K and ^{87}Rb providing a heat source.

The origin of the core is equally speculative. If the Earth formed as a solid and relatively homogeneous mass, then the subsequent segregation to its center of a dense metal phase would have been accompanied by the release of about 1.5×10^{31} joules of gravitational energy, enough to raise the Earth's internal temperature by about 1,500°C. If the initial temperatures were sufficiently low, then this stage could have been delayed until well after the main period of accretion. However, if the Earth formed as a hot body or was heated soon after formation (e.g., by the decay of short-lived radioisotopes), then segregation of the core would have occurred at the very beginning of the Earth's history. Alternatively, it is possible that the Earth formed as a layered body, an early accretion of dominantly metallic material being followed by a later addition of silicates.

P. G. HARRIS

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Cross-reference: *Meteorites—petrology*.

CORE GEOCHEMISTRY—See Vol. IVA

CRYPTOVOLCANIC STRUCTURE

The term *cryptovolcanic structure* has been used for complexes of approximately circular cross-section, superficially like volcanic pipes (necks) but containing little or no volcanic material. Limiting dimensions are not normally included in the definition but most of the bodies for which the term has been used are at least 2 km in diameter. The rocks in similar but smaller bodies, circular pipes, or bodies with irregular cross-section (even dyke-like) have been called *intrusive tuffs*. The large masses show signs of central uplift consisting of broken-up and contorted strata surrounded by a ring-shaped depression (rim syncline) that may be bounded or cut by faults. The whole structure often makes a topographical basin. Frequently a zonal form is recognized by the type and distribution of the fragmented material; marginal zones are made up of local wall rock, but central parts commonly consist of blocks of exotic material, mostly derived from deeper levels. The matrix typically consists of finely comminuted particles derived from the blocks together with such secondary minerals as sericite, chlorite, and calcite. Examples of the composition of the matrix being closely related to the dominant lithology of the blocks have been recorded (Pitcher and Read, 1952, in Bowes and Wright, 1967).

The bodies rarely occur singly and although themselves lacking in significant amounts of volcanic material, they are almost always associated in space and time with abundant volcanicity. Similarly, although their tectonic setting is typically anorogenic, they are fre-

quently distributed along the trend of fold axes.

The term was first used early this century (Branco and Fraas, 1905, in Bucher, 1963) in a description of the Steinheim basin in southern W Germany, a complex which is one of a group (Miocene in age) including the Ries basin (21–24 km in diameter), probably the biggest cryptovolcanic structure in Europe (Fig. 1). In the Ries basin there is no obvious uplift, but blocks of Permian to Jurassic strata of varying size in a matrix of ground rock-powder are scattered up to 20 km from the edge of the crater. Bodies similar in form have been described from most continents and from various geological periods. These include the Bosumtwi crater, Ghana (7 km diameter and (?) late Precambrian age), Wells Creek (13.7 km), Hicks Dome (15 km), and associated masses (Cretaceous) from Illinois, Kentucky, and W Tennessee, U.S.A., and the Wolf Creek crater and Mount Abbott mass, W Australia. Smaller bodies are even more widespread: amongst the best-documented are those from the Midland Valley of Scotland of late Carboniferous age (Francis, 1960), from the Scottish Caledonides (e.g., Bowes and Wright, 1967), from SW Cork, Eire, of late Carboniferous age (Coe, 1966) and those of Miocene age from the Urach area of southern W Germany (Cloos, 1941, in Reynolds, 1954). The last example was used as a basis for outlining the concepts of a gas-solid system having both fluid properties and erosive powers and it and other examples were cited by Reynolds (1954) in the first use of the term *fluidization* in a geological context. Cloos observed that much of the Jurassic limestone in the pipes could be matched with wall rock and had not fallen haphazardly as back-fill after eruption of explosive violence. The fluid system was shown to be highly abrasive and capable of expanding available fissures into which entrained material was transported. Reynolds extended the use of the term and suggested that fluidization-erosion is probably the chief mechanism for the formation of volcanic pipes. The principle is now established but the origin of the gas is still controversial. "Primary" gas effusion must have been an important process in the early history of the crust, and release of gas following rapid crystallization of rising magma may locally have been important, though doubts have been cast by McBirney (1963, in Coe 1966), who showed that the process envisages impossibly rapid crystallization of immense volumes of magma. McBirney's alternative suggestion, that the rise of magma to ground-water levels results in fluidization, whilst of some applicability, cannot be involved in those cases where it can be shown that the process originates at very considerable depth. Dissocia-

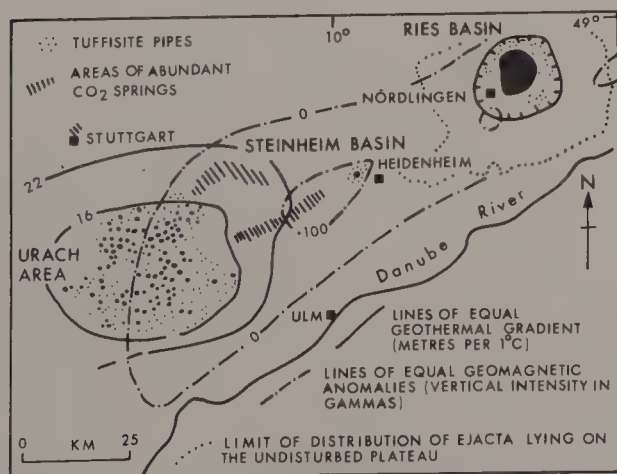


FIGURE 1. Map of the structural axis on which lie the Ries basin, Steinheim basin and the Urach area of breccia pipes, W Germany. (After Bucher, 1963.)

tion of carbonate-rich rocks by incorporation in a magma is another possible origin of gas. The direct association with magma is suggested from breccia pipes in the Avon district of Missouri, U.S.A., where deep drilling has revealed an increase in abundance of lapilli of basalt with depth. The fluidization process has also been applied to kimberlite pipes (Dawson, 1980) and to other brecciated intrusions.

Cryptovolcanic structures embrace astroblemes because, while in many cases criteria for volcanic origin are not clear, evidence for an origin by meteorite impact is at best equivocal. Thus, both "cryptovolcanic" and "impact" models have been proposed for many "astroblemes," e.g., the Ries and Steinheim basins, Germany (see various papers in Roddy et al., 1977). The term *pseudovolcanic* is also used for a crater that is possibly meteoric in origin but is more probably the result of phreatic explosion or cauldron subsidence.

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Cross-references: *Astrobleme*; *Breccia pipe*; *Calderas and volcano-tectonic depressions*; *Diatreme*; *Fluidization*; *Shock metamorphism*.

CUMULATE TEXTURES

The nomenclature of cumulate textures reflects the concept that magmas can deposit crystals in a manner analogous to sedimentary deposition in subaqueous or subaerial environments. The gravitative concentration of crystals in magma (usually by crystal sinking, but also by crystal floating) can be simulated in the laboratory, and may have been operative in

certain lava flows and minor intrusions. However, in larger intrusions of basic-ultrabasic composition, such features as low magma viscosity, high density contrast between crystals and melt, and an extensive temperature gap between initial crystallization and final solidification are conducive to the operation of crystal settling. Intrusions in which this has been an important process should typically display structural and textural features analogous to those of subaqueous sediments.

Igneous rocks formed by sedimentation are termed *cumulates* (Latin *cumulus*, a heap), and characteristically display *cumulate textures*. The general development of ideas on this subject is described by Wager and Brown (1968). However, Irvine (1982) redefines cumulate so that crystal settling is a possible, but not essential, process in the origin of the rocks to which the term is applied. He recommends that the classification of cumulate textures should be nongenetic, but retains the terminology for the textures described by Wager et al. (1960). These descriptions remain valid, but the genetic interpretation that influenced this terminology remains a matter for conjecture, largely because the role of gravity settling is critically dependent on magma viscosity and the relative densities of crystals and parental liquid, which in the case of plagioclase feldspars and basaltic liquid often remain to be demonstrated. Thus, while Wager et al. (1960) thought of the accumulative process being mainly one of crystal settling, Irvine (1982) permits this as one possibility. Other possibilities are flotation accumulation, accretion onto the margins of an intrusion, or formation of crystal "clouds" suspended in the magma because of low density contrasts between crystals and melt, high melt viscosity, or the presence of barriers in a density-stratified liquid body (see **Double-diffusive convection**).

Basis of nomenclature of cumulates

The terminology of cumulate textures used here was introduced by Wager et al. (1960), and is based on the fundamental distinction between the accumulated particles and the interstitial cementing material. This concept has long been familiar to sedimentary petrologists, but was only applied to igneous rocks as recently as 1939, in the classic account of the Skaergaard intrusion (E Greenland) by Wager and Deer (1939).

Definitions of textural terms

Cumulus (noun)—the mass of accumulated crystals before solidification of the interstitial liquid.

Cumulus (adjective)—used to describe the components of the crystal accumulation (e.g., cumulus crystals, cumulus olivines). In addition, the term *primocryst* (noun) is sometimes used (see Wager and Brown, 1968) to refer to a potential cumulus crystal, especially when discussing its nucleation and pre-accumulation behavior (e.g., plagioclase primocrysts).

Intercumulus (adjective)—used to describe the material filling the interstices of the cumulus; it may be applied either to the interstitial magma itself (e.g., intercumulus liquid) or to the phases resulting from the solidification of the interstitial magma (e.g., intercumulus material, intercumulus pyroxene). Jackson (1968) uses the term *postcumulus* instead of intercumulus when referring to the solidified interstitial material.

Cumulate (noun)—an igneous rock formed by the accumulation of crystals (originally implying gravitative setting, but compare Wager et al., 1960, with Irvine, 1982); it is the sum of the cumulus and intercumulus components.

Classification of cumulates

The classification of cumulates is based on the two textural components, cumulus and intercumulus. The cumulus mineral assemblage, which reflects the accumulation history (analogous to the factors of provenance and dispersal in the formation of clastic sediments), is indicated by prefixing the term cumulate with a list of the cumulus phases, in order of decreasing abundance (e.g., olivine cumulate, olivine-chromite cumulate, olivine-plagioclase cumulate, plagioclase-olivine-pyroxene cumulate—Fig. 3a–d).

The intercumulus material reflects the post-accumulation history of the crystal mass (analogous, perhaps, to lithification and diagenesis of clastic sediments). Various types of intercumulus crystallization have been described (Wager et al., 1960; Jackson, 1968; Wager and Brown, 1968; Irvine, 1982; Wadsworth, 1985), and these may be indicated by the use of further prefixes to the term cumulate. The following principal categories are recognized.

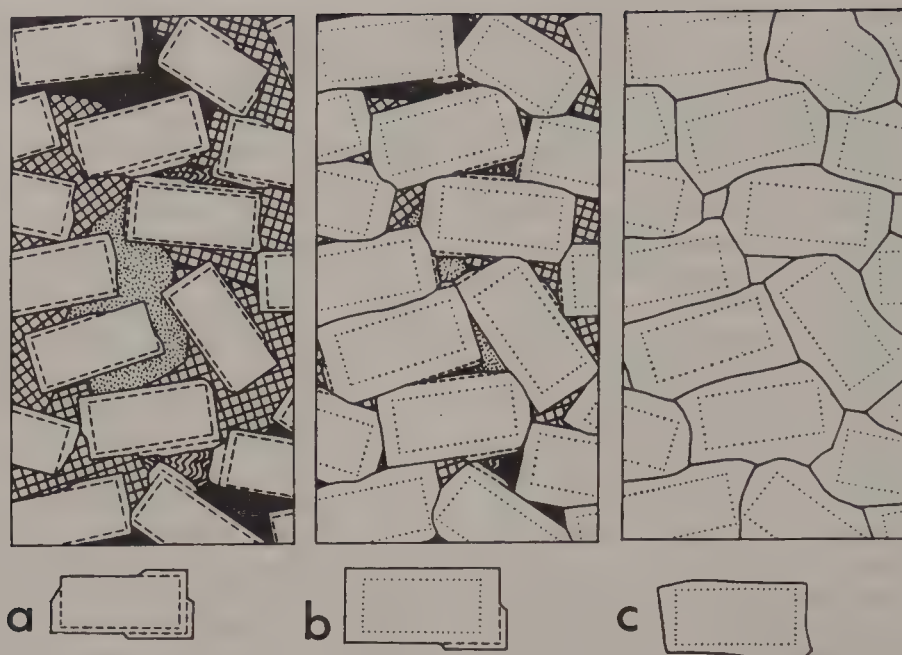
Orthocumulates. These are cumulates in which the intercumulus liquid crystallized in a straightforward fashion, precipitating progressively lower-temperature material, partly as normally zoned overgrowths to cumulus crystals (where the appropriate phase was available) and partly as new intercumulus minerals (Figs. 1a, 3e,f). Thus, in orthocumulates the bulk composition of the intercumulus material is virtually identical to that of the original intercumulus magma. This situation seems to

be relatively rare, except in the Skaergaard intrusion, and probably implies unusually fast accumulation and solidification of cumulus crystals.

Adcumulates. In certain cases it seems that lithification was achieved while the cumulus crystals lay at the margins of the mass, presumably indicating conditions of relatively slow accumulation. In such cumulates, crystallization was completed by simple overgrowths of the cumulus grains at constant temperature (i.e., without compositional zoning). This process is termed *adcumulus growth*, and the resulting rocks are *adcumulates* (Fig. 2a). In the extreme case of a single-phase cumulate (e.g., plagioclase cumulate), *adcumulus growth* could lead to the development of a monomineralic rock (Figs. 1c, 3g). Other distinctive varieties of *adcumulate* are the *heteradcumulates* (or *poikilitic adcumulates*—see Wadsworth, 1985), in which some intercumulus nucleation occurred, so that there was growth from these nuclei as well as from the adjacent cumulus crystals (Figs. 2b, 3h) and the *crescumulates* in which crystal accumulation was extremely slow or temporarily halted, so that the *adcumulus growth* not only filled the intercumulus spaces but was also able to extend upwards from the cumulus-magma interface (Fig. 2c), e.g., olivine crystals oriented almost at right angles to the layering in *harrisitic texture*. Other explanations of *adcumulus growth* involving such processes as infiltration metasomatism (Irvine, 1980), intercumulus convection (Tait et al., 1984), or mechanical compaction (Sparks et al., 1985), have also been proposed.

Mesocumulates. These are intermediate between the orthocumulates and *adcumulates*, and represent the situation in which some *adcumulus growth* took place before the intercumulus liquid was effectively sealed off by further crystal accumulation (Fig. 1b). The portion of the intercumulus material formed from the completely trapped magma is termed *pore material*, to distinguish it from the *adcumulus* component. In general, although it is not easy to make a clear distinction between orthocumulates and mesocumulates, it is obvious that some degree of *adcumulus growth* is the rule rather than the exception among magmatic sediments.

Using a combination of these two aspects of cumulate classification it is possible to give a precise and unambiguous description of any cumulate (e.g., plagioclase orthocumulate—Fig. 3e; olivine-plagioclase heteradcumulate—Fig. 3c), thus obviating the need to use more general and texturally nondescript terms such as peridotite, harzburgite, and gabbro to describe specific units in a layered intrusion.



PLAGIOCLASE : Boundary of the cumulus crystals (labradorite) diagrammatically shown by the innermost rectangle. The limits of medium and low temperature zones, where developed, shown outside the cumulus crystal boundaries.

PLAGIOCLASE : Boundary of the cumulus crystals (labradorite) shown by the dotted line. Outside is adcumulate growth of plagioclase of similar composition. In places beyond the broken lines, lower temperature zones are shown.

PLAGIOCLASE : The cumulus part of the crystal is shown within the dotted line. This has been enlarged by growth of more plagioclase of the same composition, which fills the crystal interstices.

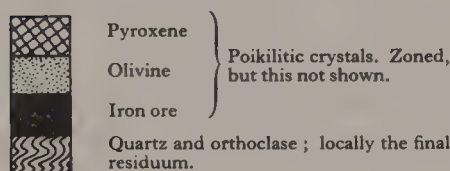


FIGURE 1. Diagrammatic representation of plagioclase cumulates formed from gabbroic magma: (a) extreme plagioclase orthocumulate, (b) plagioclase mesocumulate, (c) extreme plagioclase adcumulate. (From Wager et al., 1960.)

Description of cumulate textures

In practice, the description of cumulate textures involves the ability to distinguish between the cumulus and intercumulus components. This is relatively easy with orthocumulates, where the euhedral or subhedral outlines of the original cumulus grains should be indicated by the compositionally zoned overgrowth of these phases (since most cumulus minerals belong to solid-solution series), and the remaining intercumulus material should be obviously interstitial in habit as well as distinctive in composition (Figs. 1a, 3e, f). However, the situation is less clear among the more frequently encountered mesocumulates and adcumulates. Even here it is generally possible to recognize which minerals are present as cumulus phases, provided that at least one relatively abundant constituent is restricted to an intercumulus occurrence, since its interstitial

(and probably poikilitic) habit will contrast with the discrete (but not necessarily euhedral, or even subhedral) nature of the cumulus crystals (Fig. 3a–c). On this basis, a rock made up of only one cumulus phase will display the most striking cumulate textures (Figs. 1b, 2b, 3a), and as the number of cumulus phases increases, the texture becomes progressively less distinctive (Figs. 2a, 3c, d). Another general rule is that for a given cumulus mineral the texture becomes less distinctive with a decrease in the number of cumulus crystals per unit volume. This is because the intercumulus additions to each of these crystals will be more extensive as fewer growth centers are available, and the interstitial habit of the overgrowth may tend to mask the fact that discrete cumulus cores are present (Fig. 3c).

Another difficulty concerns the precise delimitation of the mesocumulate group, since on

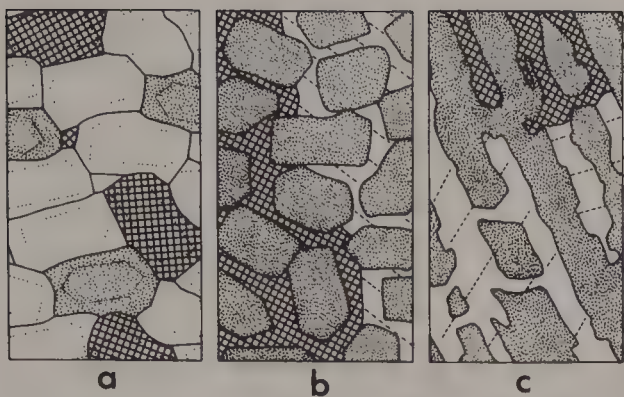


FIGURE 2. Diagrammatic representation of an adcumulate, heteradcumulate, and crescumulate formed from gabbroic magma: plagioclase, white; olivine, stippled; pyroxene, cross-hatched. (From Wager et al., 1960.) (a) Plagioclase-olivine-augite adcumulate in which the three types of cumulus crystals are considered to have been enlarged by the adcumulus process until all the intercumulus liquid has been eliminated (cf. Fig. 1c). The boundaries of the cumulus crystals of plagioclase, olivine, and pyroxene are indicated diagrammatically by dotted lines. The material inside and outside the dotted lines has the same composition. (b) Olivine heteradcumulate with large poikilitic augite and plagioclase crystals, essentially unzoned, that surround cumulus olivines: adcumulus growth of the olivines not indicated and in this diagram it is assumed that there was no trapped liquid. (c) Olivine crescumulate with all olivine showing orientation and being essentially unzoned; the surrounding plagioclase and pyroxene are also essentially unzoned and are considered to have been formed as in the heteradcumulates.

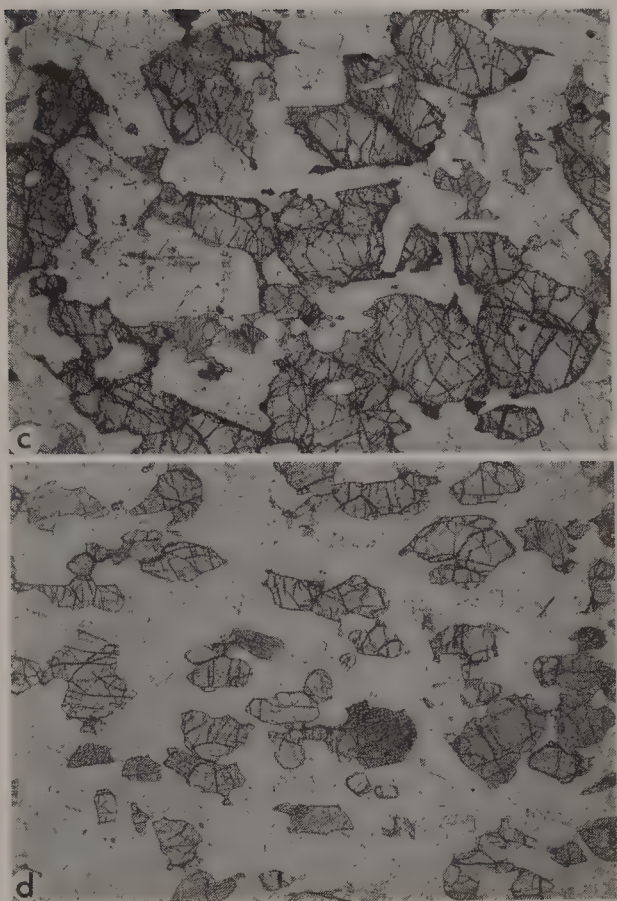
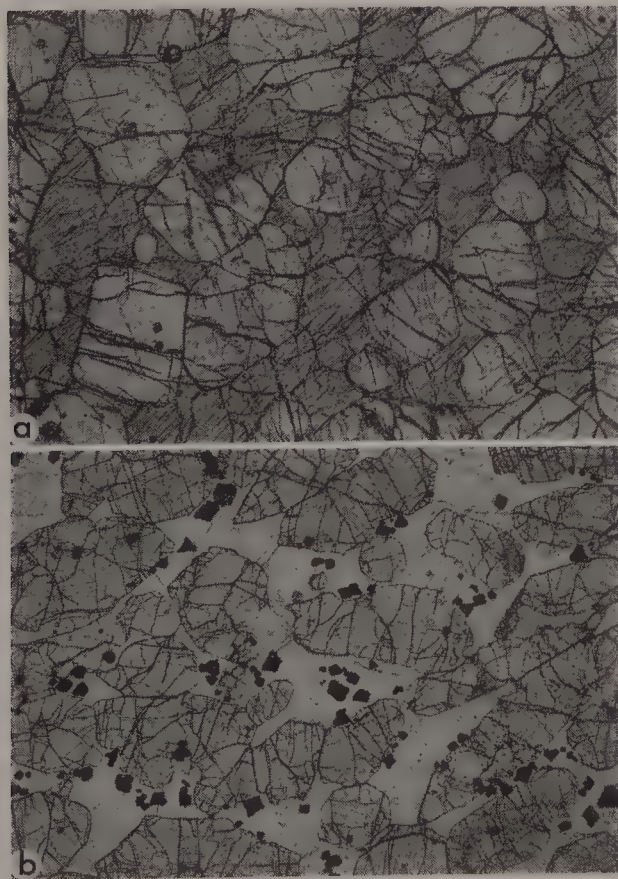


FIGURE 3. Photomicrographs of textures of cumulates; all, except (f), show the original orientation of the cumulus, which is assumed to have been deposited on a horizontal or quasi-horizontal surface. (a) Olivine cumulate with intercumulus pyroxene; PPL, $\times 12$. (b) Olivine-chromite cumulate with intercumulus plagioclase; PPL, $\times 9$. (c) Olivine-plagioclase cumulate; the relatively small number of cumulus olivine crystals has resulted in rather extensive intercumulus overgrowth of each grain, so that the original shape has been considerably modified; despite this, the discrete form of the olivines indicates their cumulus origin, and contrasts with the pyroxene which is entirely of intercumulus occurrence; PPL, $\times 9$. (d) Plagioclase-olivine-pyroxene cumulate; all the intercumulus material consists of overgrowths on the cumulus grains, and the texture is not particularly distinctive; PPL, $\times 9$. (e) Plagioclase orthocumulate, with intercumulus pyroxene, iron oxide and apatite; PPL, $\times 4$. (f) Detail of plagioclase orthocumulate, showing a plagioclase crystal with its cumulus core and zoned overgrowth, as well as intercumulus quartz; XN, $\times 20$. (g) Plagioclase adcumulate; an extreme case of a single-phase (apart from a small amount of cumulus chromite) cumulate, in which adcumulus growth has gone to completion, producing a virtually monomineralic rock; note well-developed igneous lamination; PPL, $\times 9$. (h) Plagioclase heteradcumulate; similar to (g) but local intercumulus nucleation of olivine (upper half) has allowed the development of a poikilitic, but unzoned, olivine crystal (similar in composition to cumulus olivines occurring in adjacent layers) at the same time as the adcumulus extension of the plagioclase crystals; PPL, $\times 12$.

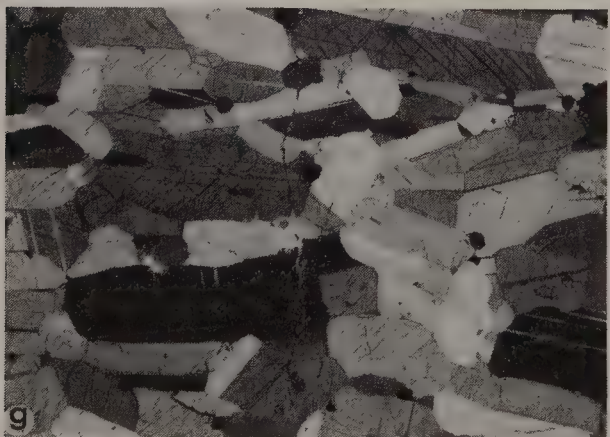
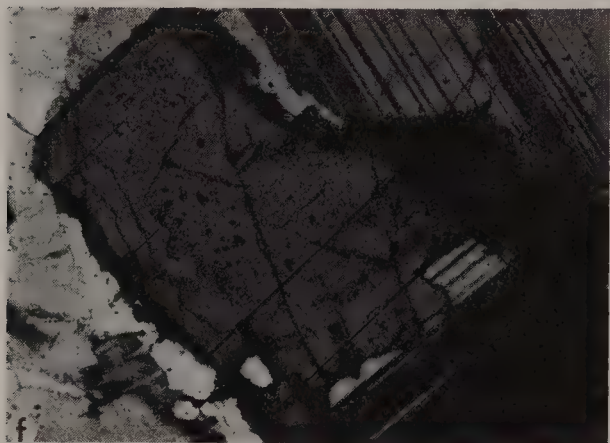


FIGURE 3. (continued)

a strict definition it is likely that most cumulates are neither extreme orthocumulates nor extreme adcumulates. In practice, in cases where there is a considerable amount of zoned material around the cumulus crystals (especially plagioclase, where the variable extinction is a particularly sensitive indicator of compositional zoning), and obvious low-temperature material such as quartz and alkali feldspar interstitially, the rock is termed an *orthocumulate*. Where there is little or no zoning (or low-temperature interstitial constituents), it is termed an *adcumulate*. These are subjective criteria and depend to a considerable extent on the original proportions of cumulus crystals to intercumulus magma. On this point, the consensus of opinion (based on sedimentation experiments and examples of extreme orthocumulates) is that the original porosity of the cumulus is rarely <20%, and more typically lies in the region of 50%. Irvine (1982) has attempted to quantify the proportions of "trapped" (low-temperature) material appropriate to the three types of cumulate, suggesting 0–7% for adcumulates, 7–25% for mesocumulates, and 25–50% for orthocumulates.

A further distinctive feature of many cumulates is the preferred orientation developed by tabular or elongate crystals coming to rest with their long axes parallel to the top of the cumulus. This type of texture is referred to as *igneous lamination* (Fig. 3g).

W. J. WADSWORTH

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* Contains many important references to earlier work.

Cross-references: *Anorthosite*; *Banding in igneous rocks*; *Bushveld Complex*; *Double-diffusive convection*; *Gabbro and norite*; *Intrusion—layered*; *Magma*; *Magmatic currents and flow*; *Magmatic differentiation*; *Nucleation of minerals in magmas*; *Skaergaard intrusion*; *Ultrabasic igneous rocks*.

D

DEPTH ZONES

During the first half of this century the idea of depth zones dominated the conceptual thinking of metamorphic petrologists. Stated simply, it indicates a primary correlation between metamorphic processes and depth. The implications are important in that, if it is correct, there is a continuous set of zones parallel to the surface of the Earth, with grade increasing downwards. Such a system would presumably be present in all the continents, although it is not clear when the zonal structure was set up, or how long it might take to reach equilibrium. Indeed, the relation of this type of metamorphism and that related more specifically to long mountain belts is not at all apparent. A predeformation age of metamorphism would imply that the present-day distribution of zones was due to deformation, which must have moved the zones upwards; a much more local and often post-tectonic distribution of zones in deformed belts is now generally accepted.

The concept of depth zones was first proposed in a coherent form by Sederholm (1891) and Becke (1903), who related metamorphic assemblages to changes in the physical conditions of metamorphism. Becke recognized that pressure and temperature, which are the two main factors in metamorphism, increase with depth, and tend to affect metamorphic reactions in opposite senses. He considered that very often metamorphic reactions tend to a decrease in volume so that an increase in pressure will enhance reaction ("volume law"); in other cases, reactions proceed in an opposite sense to that suggested by the "volume law" ($P \rightarrow Q$, Fig. 1), with lower-density products. In this latter case, Becke considered that increasing temperature had more influence than increasing pressure. This led him to postulate two zones in the crust where behavior under changing P - T conditions could be related to the two types of reactions above: (1) an upper zone consisting of rocks with mineral assemblages that resulted from reactions essentially conforming to a "volume law" ($Q \rightarrow P$, Fig. 1), for example, the metamorphism of basic rocks to produce garnet amphibolite, and (2) a lower zone in which reactions occur in a contrary sense to the "volume law"—in this region Becke thought

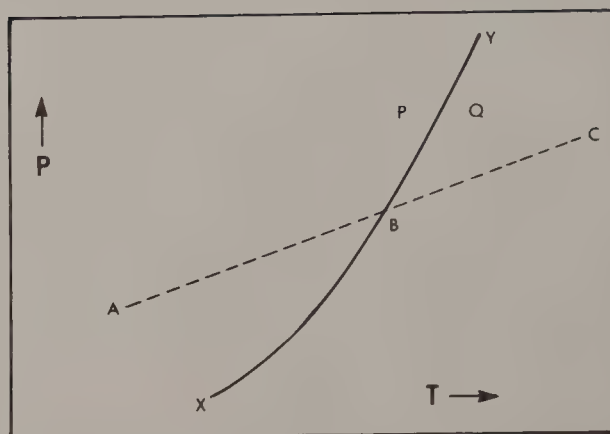


FIGURE 1. Hypothetical univariant equilibrium curve (XY) for a reaction $P \rightarrow Q$; ABC is a hypothetical gradient related to depth. (After Turner, 1981.)

that increasing temperature was the dominant factor. In this Becke was clearly wrong, for many reactions in the upper zone are dehydration reactions and take place contrary to the "volume law." Present-day thought (Turner, 1981) would explain these reactions in terms of a simple univariant equilibrium curve (XY, Fig. 1) with a positive slope for a given reaction, and for such reactions the entropy and volume increase or decrease together.

Later work, particularly of Grubenmann and Niggli (1924) emphasized and extended Becke's original ideas, increasing the number of depth zones and making a formal correlation between temperature, hydrostatic pressure, and stress (Table 1). Like earlier workers, Grubenmann was particularly impressed by the mineralogical and textural differences between cover and basement rocks, which he thought were purely a function of depth; and indeed his depth criteria were based entirely on the mineralogy and textures in the rocks.

Grubenmann and Niggli (1924) list characteristic minerals for each zone: epizonal minerals include sericite, paragonite, glaucophane, chloritoid, actinolite; mesozonal minerals include biotite, muscovite, staurolite, hornblende, kyanite; katazonal minerals include cordierite, sillimanite, spinel, wollastonite. Characteristic structures for each zone are:

TABLE 1. Depth zones; after Grubenmann and Niggli (1924)

	Temperature	Hydrostatic pressure	Stress
Epizone (upper zone)	Moderate	Mostly low	Often strong; also missing
Mesozone (middle zone)	Higher	Mostly higher	Often still very strong; also missing
Katazone (lower zone)	High	In many cases very high	Often completely missing

epizone—slaty, finely schistose; mesozone—strong mineral schistosity and lineated; katazone—granulitic or foliated and migmatitic.

Correlations such as that given in Table 1 are clearly wrong on a number of counts: (1) although some Archaean rocks belong to the katazone, many have mesozonal or even epizonal assemblages (Eskola, 1939, p. 340); (2) many metamorphic areas show *P–T* gradients that are essentially isobaric (Turner, 1981, p. 452); and (3) high-pressure minerals such as glaucophane and lawsonite clearly do not belong to the epizone. Indeed, the classic Barrovian terrane in Scotland, on the basis of index minerals, contains all three zones, although there is little evidence of a major change in depth across the Barrovian zones.

Strong reaction to the depth zone concept came from Eskola (1939) and Harker in the 1920s and 1930s, but in spite of that the idea has been entrenched in European thinking until very recently. Even now there is some residue of thinking that depth may be equated directly with grade.

M. P. ATHERTON

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Cross-references: *Metamorphic grade, metamorphic zone; Metamorphism; Metamorphism—controls.*

DESERPENTINIZATION

Deserpentinization is the process whereby serpentinite is progressively dehydrated during prograde metamorphism, ultimately leading to the reappearance of anhydrous ferromagnesian minerals. Unlike serpentinization, which involves the replacement of high-temperature phases by low-temperature hydrous phases in one step, deserpentinization involves a series of progressive steps, each marked by different assemblages roughly equivalent to the facies subdivisions of regional or contact metamorphism (Evans, 1977).

Prograde greenschist facies metamorphism usually involves the progressive replacement of chrysotile + lizardite assemblages by antigorite + lizardite, with the appearance of Mg-chlorite when spinel reacts with chrysotile during breakdown (Bliss and McLean, 1975). Talc also appears; both reactions involve the dehydration of serpentine minerals.

Talc ± serpentine ± chlorite assemblages represent the low-temperature amphibolite facies, whilst at the higher temperature end the final steps in dehydration involve the appearance of olivine and enstatite. Alongside this there is a progressive change in spinel composition, and above ca. 600°C brucite and magnesite are replaced by periclase.

The progressive introduction of CO₂ to replace H₂O may also promote dehydration. CO₂-rich fluids both permit the replacement of brucite by magnesite and alter the stability fields for some of the silicate assemblages (Fig. 1). Magnesite + talc represents most of the amphibolite facies, and the 500°C threshold at which olivine reappears, while not changing position, becomes a more complex reaction. High *P*_{CO₂} appears to promote the appearance of anthophyllite between the breakdown of talc and the appearance of enstatite and olivine (Fig. 1). Progressively higher *P*_{CO₂} lowers the “anthophyllite in” reaction temperatures (see Johannes, 1969).

The presence of calcic phases in the protolith causes some complications, leading to more com-

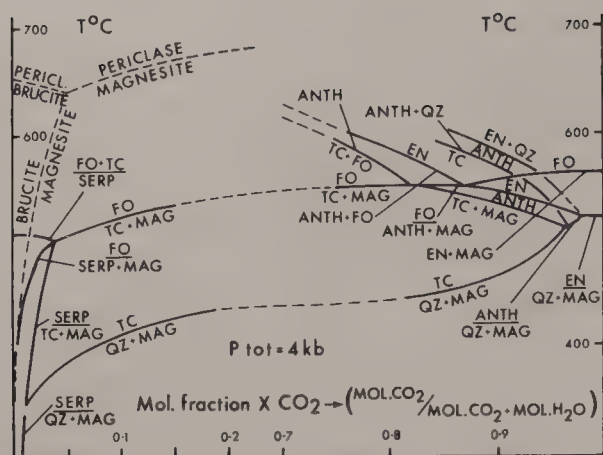


FIGURE 1. T - X_{CO_2} diagram for serpentine phase relations. (After Park, 1983.)

plex phase assemblages. Dolomite and calcite become important, often replacing magnesite, while calcic amphiboles, mainly tremolite or actinolite, more rarely cummingtonite, represent the amphibolite facies. Final dehydration in these cases results in the formation of diopside.

ADRIAN F. PARK

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Cross-references: *Amphibolite facies*; *Contact (thermal) metamorphism*; *Greenschist facies*; *Metamorphic facies series*; *Regional metamorphism*; *Serpentinite*; *Serpentinization*.

DIABASE

The term *diabase* was originally applied by A. Brongniart in 1807 for rocks later recognized as diorite. Some authors use it to refer to an intrusive rock, commonly occurring as dykes and sills, whose main constituents are labradorite and pyroxene and whose texture is characteristically ophitic or subophitic. This

usage is synonymous with that of dolerite. Other authors use the term diabase for pre-Tertiary dolerites; as these rocks are frequently altered, this usage has come to refer to altered dolerite in which feldspars are saussuritized or albitized and the pyroxenes show replacement by amphibole or chlorite.

Diabasic has been used as synonym of ophitic. Kemp (1900) considered that diabasic applied to those textures in which there was a predominance of plagioclase with augite in the interstices, while ophitic indicated a predominance of augite over plagioclase.

The following terms incorporating diabase are used.

Ashbed diabase—a rock of a scoriaceous, amygdaloidal, tabular intrusion that has incorporated sandy material.

Hunne diabase—a type of quartz dolerite (diabase) containing hornblende, biotite and often chlorite in addition to plagioclase and a clinopyroxene (Hunnellberg, Sweden).

Kinne diabase—a type of olivine diabase with chlorite and secondary quartz.

Konga diabase—a type of granodolerite composed of labradorite, clino- and orthopyroxene in a microgranitic groundmass of quartz and orthoclase (Konga, Sweden).

Oje diabase—a type of porphyritic dolerite containing plagioclase laths in an aphanitic groundmass.

D. R. BOWES

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- Cross-references: *Dolerite*; *Dyke (dike)*; *Sill*; *Textures of igneous rocks*.

DIAPIRISM

The term *diapir* (from the Greek διαπείρω—to pierce) was introduced by Mrazec (1915) for intrusive salt bodies that developed from salt layers in the crests of anticlines. The term is used in a general sense for a plug-like mass of any rock that pierced through overlying rock units and became lodged in them. The process of piercing of the rock mass is called *diapirism*.

Rock types found in diapirs include salt, gypsum, serpentinite, argillaceous and carbonate rocks, as well as rocks that, during the act of piercing, were a magma, a crystal mush, or a mobilized magmatic or metamorphic rock. Salt

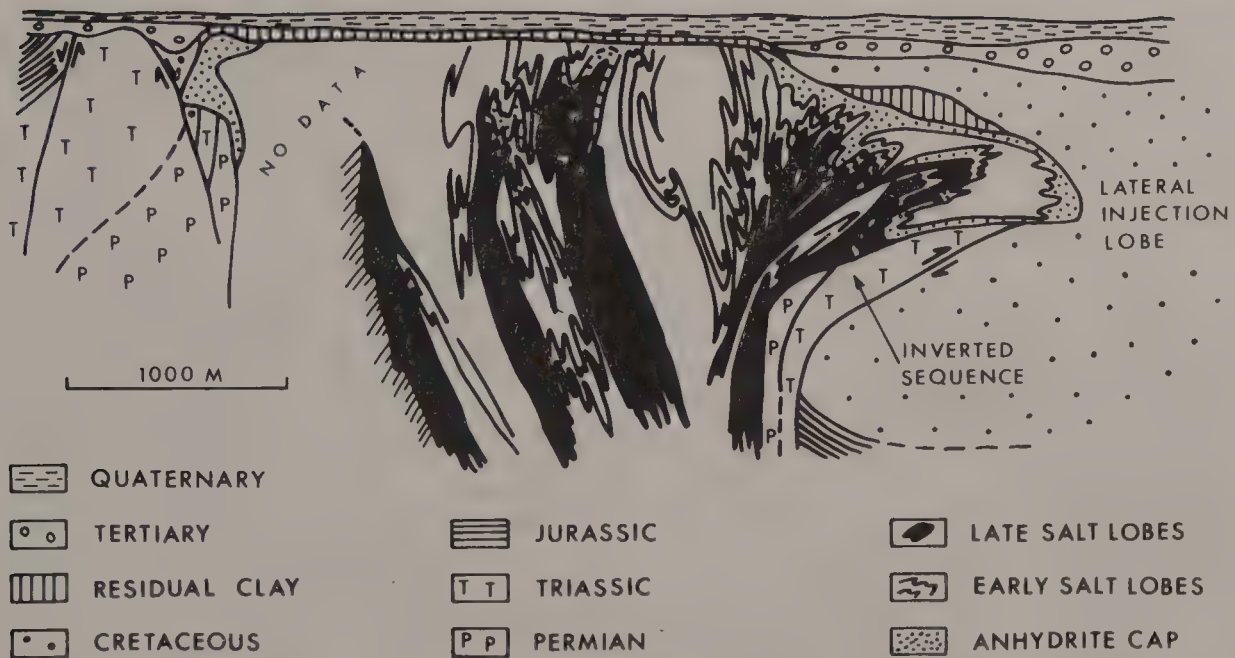


FIGURE 1. Multiple intrusion of salt diapirs in the Heide salt plug, Hanover, FDR. (After Hills, 1972.)

diapirs, such as those in northwestern Europe (Fig. 1) and on the Gulf Coast of North America, are structurally well known because of economically important deposits of salt, potash, gypsum, sulfur, gas, and oil. Their internal structure is complex (Balk, 1949). An argillaceous diapir with a polymict breccia consisting of shaly, calcareous, and dolomitic rock fragments is described by Coats (1964); other examples of diapirs in sedimentary rocks and an extensive bibliography are given by Braunstein and O'Brien (1968).

Hans Cloos (1921) initiated the detailed structural investigation of igneous diapirs, such as granitic batholiths and stocks, and volcanic plugs. A compilation of structural features resulting from the diapiric action of magmatic masses (emplacement of granites) is given by Balk (1937) and Marsh (1982) discusses the mechanisms of igneous diapirism (see **Igneous intrusions—structural behavior**).

Diapirism of mobilized rock masses in high-grade metamorphic terranes, such as deeply eroded Precambrian regions, may have contributed to the blob-like geological map pattern that characterizes these regions. These patterns may represent all gradations from doubly-plunging anticlines, through domal structures such as mantled gneiss domes, to diapirs of metamorphic rock masses, though their interpretation is fraught with controversy (cf., Snowdon and Bickle, 1976). The diapiric model for the origin of Archaean batholiths (Macgregor, 1951) has been supported by some workers who suggest that the domes had the properties of magmas and by others who considered that the batholiths moved as solid-state diapirs or

gneiss domes. Other workers view vertical movements on a batholithic scale in response to gravitational instability as being of no real significance, although individual plutons within the batholiths may have been intruded as diapirs (Snowdon, 1984, and references therein). Hickman (1984) interprets Archaean diapirism in Western Australia as involving solid-state diapirism of a granitic substratum through a denser volcanic-sedimentary layer with the resultant granitic gneiss domes not being the result of either magmatic diapirism or superimposed folding.

The diapiric mechanism is controlled in the first place by the presence of a layer or zone in which the material has a lower density than the overlying rock units. A second requirement is that the material of lower density has a high plasticity, that is, an ability to flow. And thirdly, a disturbance of the gravitational instability is needed to trigger the beginning of flow. The disturbance, which may be orogenic tectonism or another process such as erosion, causes differences in the pressure distribution of the overlying rock units. Diapirism, then, will favor those places where the pressure of the overburden on the material of lower density is least. The driving force of diapirism is the density inversion. Diapirism is, therefore, a kind of gravitational tectonism; the lighter material moves upward in intrusive bodies, while the heavier country rock moves down (Ramberg, 1967).

The pattern of minor structures reflects the diapiric mechanism. There are internal structures, formed by flow in the intruding mass, and external structures, expressing the deformation



FIGURE 2. Foliation around and within diapiric batholiths, Sierra Moderna, Spain. (After Brun and Pons, 1981.)

of the country rock by the piercing action of the diapir (Schwerdtner, 1982). A flow structure develops by rotation of elongate and platy minerals and rock fragments in the line and plane of extension of the intruding mass. Thus, parallelism of minerals and inclusions in a granite diapir represent a linear and planar flow structure that is a reflection of the pattern of flow in partly crystallized magma (Fig. 2). The deformational pattern displayed in salt diapirs is of a different nature. The structure represents the originally horizontal layering of salt deposits that became intensely distorted by rock flow. Folds in the salt layers are, in effect, doubly plunging, with fold axes bent to such an extreme that they are like vertical hairpins, and the folds are like thimbles. Less conspicuously, there is also a flow structure present, generally at an angle to the layering. The flow structure, expressed by parallelism of elongate crystals of anhydrite and halite, is due to rotation of the minerals in the direction of extension of the salt mass, which is similar in all respects to the development of flow structure in granite diapirs.

External diapiric structures include a number of different features that record the forceful invasion of material in space already occupied by rock (Fig. 3). Layering or foliation in rock adjacent to the diapir is commonly dragged upward by rotation of its plane into the direction of diapiric movement. The country

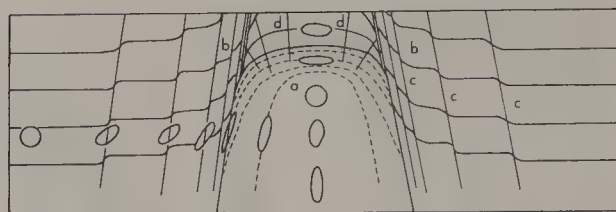


FIGURE 3. Idealized section of a diapir showing the strain ellipsoid in relation to principal structures caused by diapirism: a, flow structure; b, drag of layering or foliation; c, fault; d, extension joints in domed roof of the diapir. (After de Waard, 1949.)

rock is further displaced upward by faulting parallel to the walls of the diapir. Low-angle thrusts indicate outward displacement of country rock caused by the expanding diapir. In the roof of a diapir doming of overlying strata occurs, and collapse structures may result from extension in rocks of the dome. In the roof of magmatic diapirs, extension joints are commonly filled with magmatic rock to form dyke swarms (cf., Schwerdtner et al., 1978).

The overall fabric of spatial relationships of the internal and external structures formed by the process of diapirism is that of axial symmetry; the orientation of the axis indicates the direction of tectonic transport.

D. DE WAARD

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Cross-references: *Batholith*; *Emplacement—modes*; *Igneous intrusions—structural behavior*.

DIATREME

Diatremes are cylindrical or irregularly-shaped bodies drilled through enclosing rocks by the action of gas-charged magma (Fig. 1).

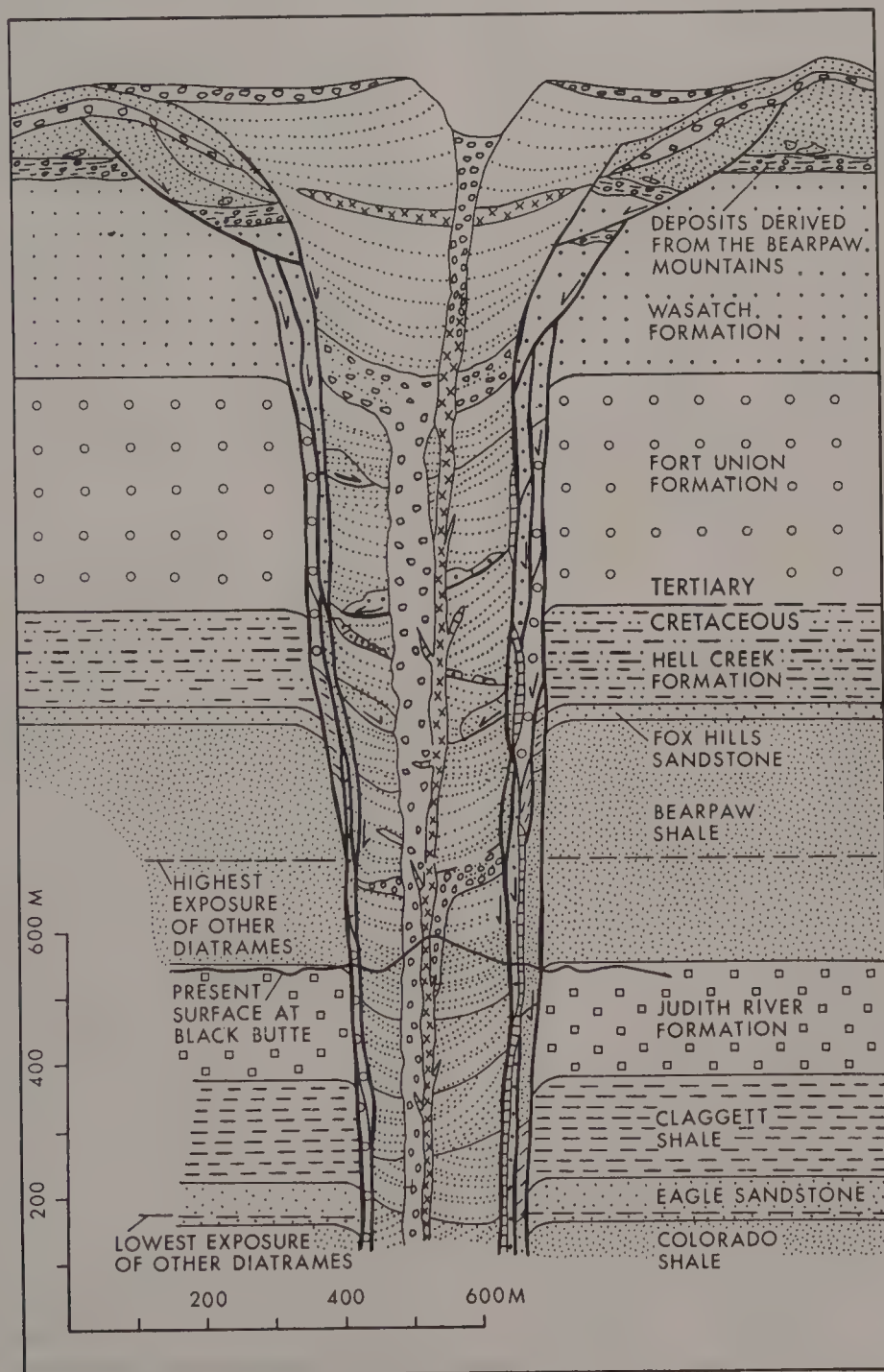


FIGURE 1. Diagrammatic section of Black Butte diatreme, Montana, U.S.A. (After Hearn, 1968.)

They are usually associated with explosive volcanic activity at the surface. Three types of structure are usually called diatremes, which probably only differ from each other in depth of formation. Explosion-breccia pipes and intrusive breccia pipes are of deep-seated origin and do not necessarily reach the surface. Tuffisitic breccia is largely a near-surface intrusive phenomenon and some tuff-filled plugs represent the surface expression of this activity.

Breccia pipes are the normal deep-seated expression of diatremes. They are subcylindrical bodies of angular brecciated country rocks in a matrix of comminuted country rock. These form in situ and shattering is found at their margins (Sage, 1978). The gases causing the explosions are derived from rising volatile-rich magmas below, which may afterwards mobilize the breccias by further evolution of gas that transports the breccia upwards forming very rounded boulders in a breccia that is now intrusive. In other cases the mobilization may be by the actual magma itself. Kimberlite pipes are a form of intrusive breccia in which the magma has acted as the mobilizing agent and in this case the diatreme is largely igneous rock. Intrusive breccia pipes commonly contain boulders that have been transported upward many hundreds of meters. This has been shown from stratigraphic considerations in the diatremes of the Navajo–Hopi region, Arizona,

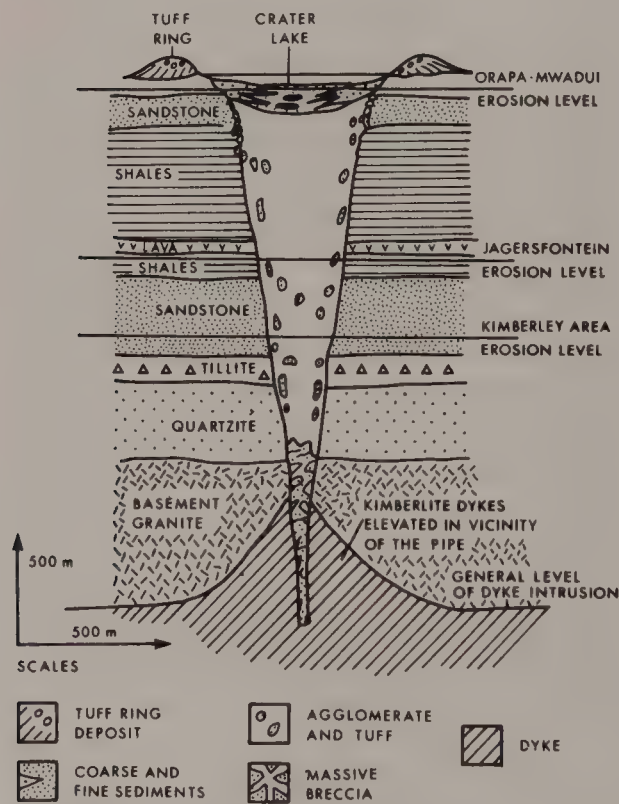


FIGURE 2. Diagrammatic section of a typical kimberlite diatreme. (After Dawson, 1980.)

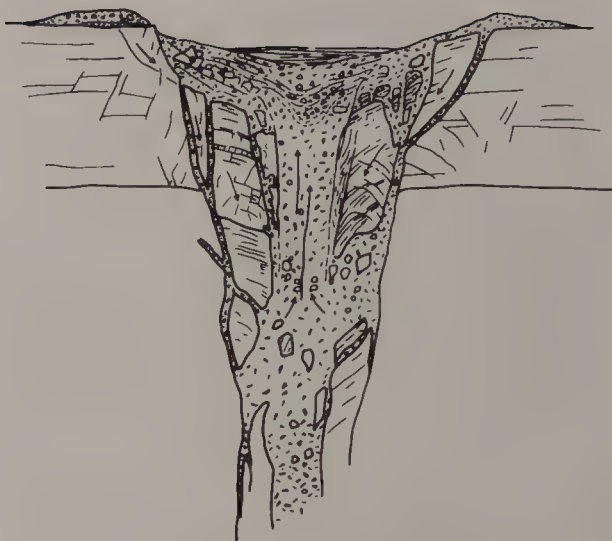


FIGURE 3. Breccia pipe with collapse downwards of large blocks of country rock and upward intrusion of tuffisite; reconstruction of Swabian tuff-pipes. (After Cloos, 1941.)

and the Eureka District, Utah, U.S.A., and in the appinite–syenite–breccia pipes of W Scotland and Eire. Kimberlites also contain abundant inclusions of peridotite and eclogite that are thought to be derived from the mantle and in these cases the upward transport would seem to be of the order of many kilometers rather than hundreds of meters (Fig. 2).

The higher level diatremes are caused when the gas-charged magma forces its way through the rock to the surface. The speed at which this intrusion takes place is such that the magma vesiculates and is supercooled to a tuff composed of glass shards, existing phenocrysts, and breccia fragments, all of which are transported by streaming gas. This dense “sand blast” rounds the breccia fragments and causes the intrusion of the tuffaceous material into cracks in the wall-rock. Fragments are broken off and rise up the expanding pipe. Larger fragments subside into the pipe. The tuff may be injected for long distances into the wall-rock. This forms a tuffisitic volcanic neck (Fig. 3). Cloos (1941) has described this type of volcanic emplacement for the Swabian pipes, W Germany, and similar series of volcanic pipes have been described from Fife, Scotland (Francis, 1962), the Marysville Buttes, California (Williams, 1929), and the Navajo–Hopi Buttes, Arizona, U.S.A.

At this high level in the crust there is little doubt that the intrusion, the break up of the wall-rock, and the vesiculation of the magma were caused by the breaching of the surface by a volcanic explosion, the consequent lowering of the pressure below causing an extremely rapid exsolution and ascent of gases from the highly



FIGURE 4. Rhythmic tremolite-actinolite overgrowths on corroded pargasitic hornblende which has a core of clinopyroxene, itself with a corroded margin; Back Settlement, Scotland, XN, $\times 32$. (See Bowes et al., 1964.)

volatile magmas. This process has been termed fluidization (Reynolds, 1954) and is probably the major form of emplacement of high-level diatremes. Intrusive breccias such as those associated with kimberlites, the appinite suite of the Caledonides of W Scotland, the latites and quartz monzonites of the Rocky Mountains, and carbonatite volcanoes, are probably emplaced in a similar manner, without necessarily reaching the surface (Wright and Bowes, 1968).

The formation of explosion-breccia and the mechanism of brecciation of the rocks that are later mobilized to form intrusive breccias is of less certain origin. Rapid variations in the crystallization conditions of appinite associated with explosion-breccias in W Scotland leads to a series of tremolite-actinolite overgrowths on the hornblende crystals (Fig. 4). This, together with the resorption of the hornblende crystals, is the only evidence so far established of the rapidly changing crystallization conditions that would be associated with the build-up and release of gas pressure during a series of explosions (see *Appinite*).

Phreatic water is not essential to the formation of diatremes, for, although all the largest surface volcanic eruptions appear to be caused by the influx of large amounts of sea water or groundwater, there is a much closer

relation between diatremes and the volatile content of magmas, the rock type, the stage of differentiation, and the presence of breccias, than with other superficial factors.

A. E. WRIGHT

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Cross-references: *Appinite*; *Breccia pipe*; *Carbonatite*; *Explosive volcanic eruptions—classification*; *Fluidization*; *Kimberlite*; *Tuffsite*; *Volatiles*.

DIFFERENTIATION INDEX

Laboratory studies of silicate melts and igneous rocks indicate that most residual melts produced by fractional crystallization become progressively enriched in silica and/or alkali aluminosilicates. These constituents all lie in the system $\text{SiO}_2\text{--NaAlSiO}_4\text{--KAlSiO}_4$ (Fig. 1). This system is the goal toward which these residual liquids move, but cannot pass, and for this reason it was called "petrogeny's residua system" by N. L. Bowen. The residua character of this system can be illustrated by two examples. Figure 2 depicts relations in the system $\text{Mg}_2\text{SiO}_4\text{--KAlSi}_2\text{O}_6\text{--SiO}_2$, as shown by laboratory studies on synthetic melts; here crystallization gives rise to a residual liquid that with extreme fractionation consists of 98% of the components of petrogeny's residua system. Figure 3 shows the relationship in a number of porphyritic volcanic rocks between the content

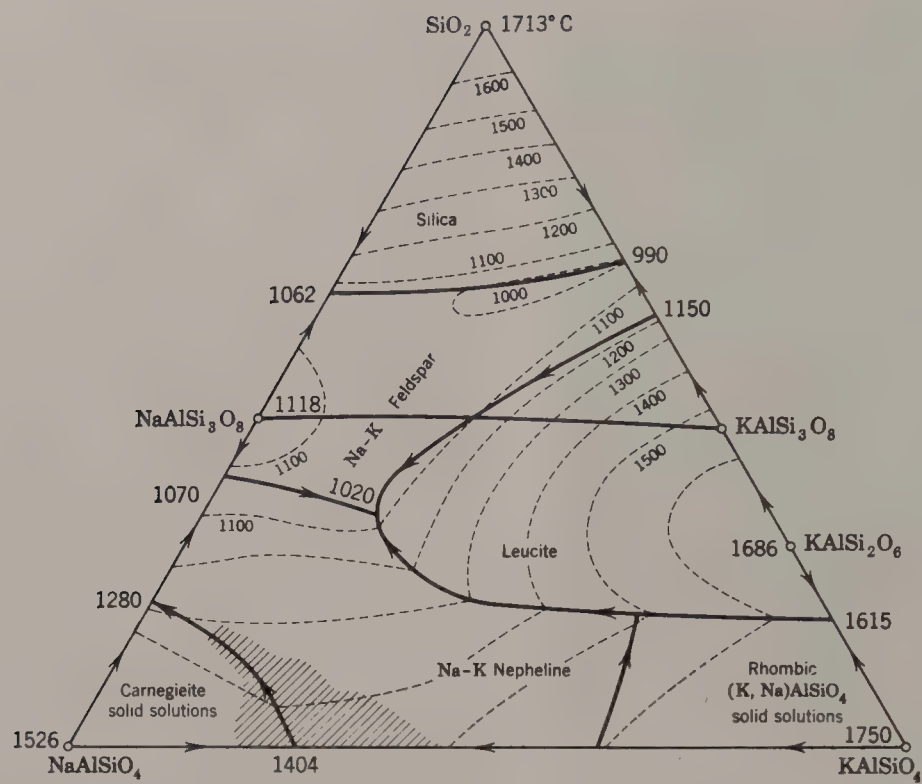


FIGURE 1. The ternary system NaAlSiO₄-KAlSiO₄-SiO₂; arrows indicate the directions of falling temperature; primary phases are shown and their corresponding fields of crystallization are outlined. Natural nephelines are not pure NaAlSiO₄, but include both SiO₂ and KAlSiO₄ in their composition. (From Barth, 1962.)

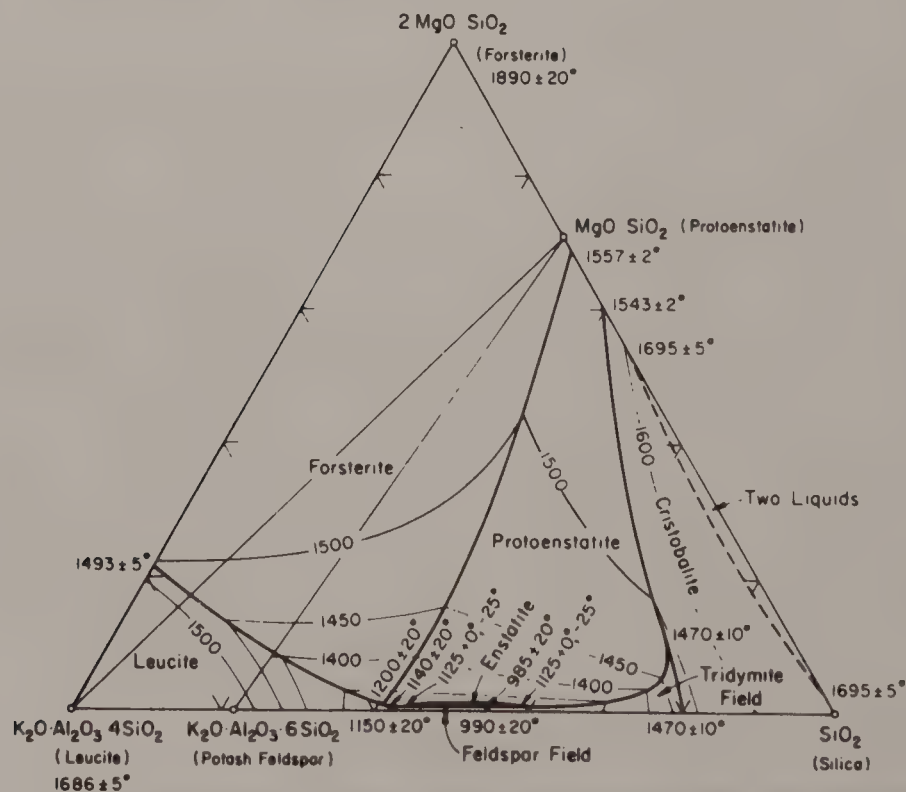


FIGURE 2. Phase equilibrium diagram to illustrate the pronounced enrichment of the liquid in orthoclase on crystallization in the system Mg₂SiO₄-KAlSi₂O₆-SiO₂. (From Thornton and Tuttle, 1960.)

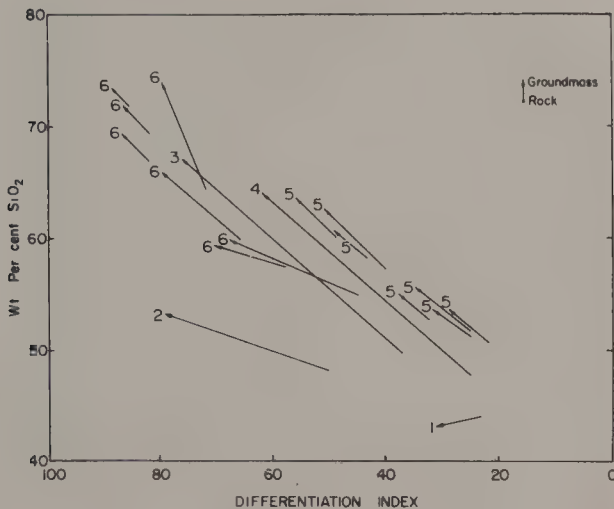


FIGURE 3. Diagram illustrating the enrichment of the groundmass of some porphyritic rocks in the constituents of petrogeny's residua system: 1, Katmai region, Alaska; 2, Vesuvius; 3, Kinkell, Scotland; 4, Kap Doussy, E Greenland; 5, Hakone volcano, Japan; 6, San Juan region, Colorado, U.S.A. (From Thornton and Tuttle, 1960.)

of petrogeny's residua in the whole rock and in the aphanitic or glassy groundmass. Such a groundmass represents the residual liquid produced by crystallization of the phenocrysts and is, without exception, richer in the constituents of petrogeny's residua system than is the rock as a whole.

A liquid rich in the components of petrogeny's residua system can be formed by processes other than fractional crystallization. For example, the early liquids formed by partial fusion of almost any silicate rock will consist almost entirely of the components of this system.

If petrogeny's residua system is considered as the base of a tetrahedron whose fourth apex represents all the other constituents of a rock, the distance from that apex to the point representing that rock is a measure of the richness of that rock in petrogeny's residua system. This distance, called the *differentiation index* (DI), is the sum of the percentages of quartz, orthoclase, albite, nepheline, leucite, and kaliophilite that occur in the norm of the rock.

However, intrusive basaltic (in particular, tholeiitic) melts may also show a second line of liquid descent that, at least in its early stages, may be more significant than enrichment in petrogeny's residua system; this is the enrichment of the remaining melt in Fe. Such a change in composition of the remaining melt would produce no change in the differentiation index, and for this reason von Gruenewaldt (1973) has proposed a *modified differentiation index* (MDI). This is defined as the sum of the

quartz, orthoclase, albite, nepheline, leucite, kaliophilite, hedenbergite, ferrosilite, and fayalite in the norm of the rock recalculated as free of magnetite, ilmenite, apatite, and other accessory minerals.

The differentiation index has proved to be an ideal quantity for illustrating variations in the chemistry of the igneous rocks, for the range in differentiation index between the most acidic and the most basic rock is the same (0–100) in silica-undersaturated, silica-saturated, and silica-oversaturated series. It can be used to show both the average variation in composition for all igneous rocks and the variation within the suites of rocks from individual areas.

Frequency diagrams of oxide percent vs. differentiation index show the average variation (cf., Thornton & Tuttle, 1960). On the basis of these diagrams it can be seen that as the differentiation index increases (1) SiO_2 and K_2O increase continuously in the majority of the analyses, (2) CaO , MgO , and FeO decrease continuously, (3) there is little or no change in Na_2O and Fe_2O_3 , and (4) the SiO_2 and K_2O diagrams show two maxima corresponding, presumably, to the granites and rhyolites on the one hand and to the gabbros and basalts on the other.

Similar diagrams can be used to illustrate the variations in composition of rocks from a given area. Figure 4 shows diagrams for SiO_2 and

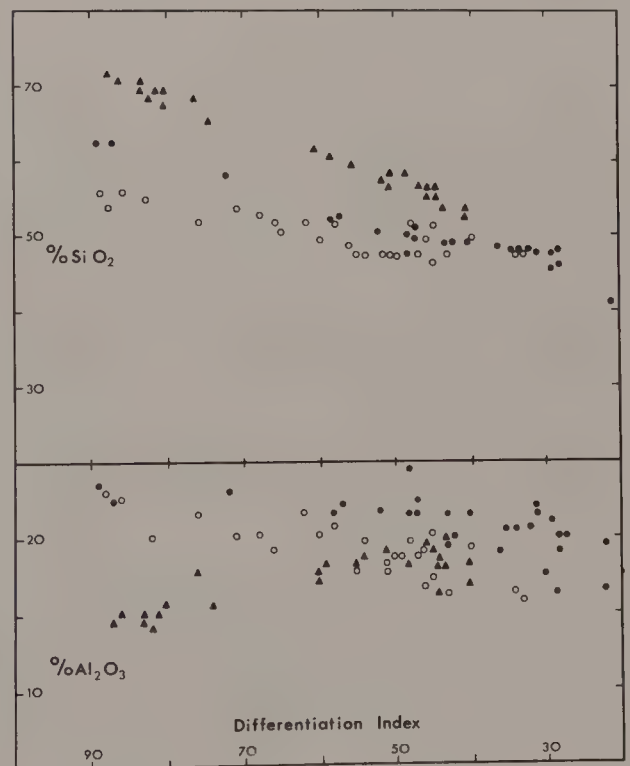


FIGURE 4. Oxide vs. differentiation index diagrams: ○, Vesuvius (undersaturated); ●, Kohala, Hualalai, and Mauna Kea, Hawaii (saturated); ▲, Crater Lake, Oregon, U.S.A. (oversaturated).

Al₂O₃ vs. DI for three rock suites: one silica-undersaturated, one silica-saturated, and one silica-oversaturated. It should be noted that in a particular suite, if the variation is due to fractional crystallization, the degree of differentiation of a specific rock is indicated not by its differentiation index but by the difference between its differentiation index and that of the parent magma.

The differentiation index also can be used in the classification of the igneous rocks. There is only one possible differentiation index for each chemical composition and, when used in conjunction with a modal analysis, it provides evidence on the basicity of the rock that the modal analysis alone does not supply. Each of the various rock types in a modal classification has a wide range in chemical composition because of the extensive solid solution possible in the minerals as well as of the range permitted by variations in the amounts of minerals present. The use of the differentiation index in addition to the modal classification gives the petrologist some knowledge as to where a given rock falls in the range of admissible compositions.

CHARLES P. THORNTON

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Cross-references: *Experimental petrology*; *Magmatic differentiation*; *Petrochemical calculations*; *Reaction principle and reaction series*.

DIFFUSION, GEOLOGICAL ROLE—
See Vol. IVA

DIORITE AND TONALITE

Definition

Diorites are holocrystalline coarse- to medium-grained intermediate plutonic igneous

rocks composed of plagioclase (An_{<50}, generally 50–75 vol.% of total) and mafic constituents (often hornblende, but also biotite or augite). Leuco- and meladiorites also occur. Minor amounts of alkali feldspar and quartz may be present. With increasing quartz they grade into *quartz diorite* (Q/Q + A + P 5–20%) and *tonalite* (Q > 20%), and with increasing alkali feldspar (A/A + P > 10%) into *monzodiorites* (often called *syenodiorites*) and quartz monzodiorites. Tonalites grade into *granodiorites* with increase in alkali feldspar content. Some authors refer to leucocratic tonalite or granodiorite containing very little or no alkali feldspar as *trondhjemite*. Feldspathoid-bearing variants occur, but are comparatively rare. With more calcic plagioclase, diorites grade into gabbros. The name diorite is from the Greek “to distinguish” and is due to R. J. Haüy; tonalite is named after Mount Tonale, Tyrol, by G. von Rath (see Johannsen, 1937).

Mineralogy

The usual composition of the plagioclase is andesine. It is often zoned from labradorite with an albitic outer rim. The usual essential colored minerals are green or brown hornblende, sometimes with an outer rim of pale bluish Fe-rich amphibole, biotite, diopsidic augite, and less often hypersthene. Quartz and alkali feldspar (usually orthoclase) are accessory minerals in most diorites, often as myrmekite. In diorites of syenitic affinities the plagioclase may be rimmed by microperthitic alkali feldspar of ternary bulk composition. In this association, ferro-augite, Fe-rich biotite, and sometimes olivine are common mafic constituents. Other accessories are sphene, ilmenite, magnetite, apatite, allanite, and calcite. Epidote is a characteristic, possibly deuteric, mineral in many diorites and quartz diorites.

Distribution and genesis

The calc-alkaline association gabbro–diorite–tonalite–granodiorite–granite is the dominant plutonic igneous association in the continental crust. In tectonically active regions, this association makes up the great composite “granite” batholiths; relative volumes of each type for the Central Peruvian Batholith (Cobbing and Pitcher, 1972; Pitcher, 1978) and the Southern California Batholith (Larsen, 1948) are:

	Gabbro- diorite	Tonalite	Granodiorite and adamellite	Granite
Peru	15.9	57.9	25.6	0.6
California	14	50	34	2.5

Despite sharp internal intrusive contacts in these composite plutons and the common evidence for late-stage K-metasomatism, there seems every reason to interpret (as do Cobbing and Pitcher, 1972) the sequence as representing a progressively evolved magma series. In principle such a series could be produced by fractional crystallization of dioritic magma, with the associated gabbros representing accumulative rocks, and the granites residual members. The gabbros however, are often intruded first (Cobbing and Pitcher, 1972) and these authors support a prevalent view that the main magma bodies are produced by remelting of the lower crust with subsequent diversification. The intermediate members of the sequence frequently contain fine-grained basic inclusions that are consistent with a refractory residue left after partial melting (see Piwinskii, 1968). Pitcher (1978) therefore suggests that gabbroic magma was formed prior to crustal melting and may, in part at least, have been a heat source for subsequent partial melting of crustal rocks.

Many modern models for genesis of calc-alkaline orogenic magma series involve components contributed by both subducted crust and by a mantle wedge above the downgoing slab, modified by alkalis and fluids emanating from the subducted rocks, which, in general, are wet ocean-floor basalts and sediments (Atherton et al., 1979). However, $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the Andean batholithic rocks indicate a largely mantle origin. Chappell and White (1974) have classified granitoids into I and S types, with igneous and sedimentary source rocks, respectively; silica-poor rocks, such as tonalite and diorite, are characteristically members of I-type granitic sequences.

Alkali diorites (grading into syenodiorites) may be derivatives of gabbroic bodies of alkali basalt parentage or occur as minor members of dominantly syenitic intrusions. In such complexes they may occur as distinct intrusive phases or they may show a complete gradational relationship into syenite.

Other types

Anchorite—a diorite with isolated patches of minerals and felsic veins (Anchor Inn, Nuneaton, England).

Anden-diorite—a quartz-bearing diorite with augite as its principal mafic mineral (Andes).

Belugite—the term for a group of rocks intermediate, with regard to feldspar content (i.e., containing andesine and/or labradorite), between diorite and gabbro (Beluge River, Alaska, U.S.A.); *aleutite* is a porphyritic variety (Aleutian Is.).

Chlorophyre—a green porphyritic quartz diorite.

Dungannonite—an alkalic corundum- and scapolite-bearing diorite (Dungannon, Ontario, Canada).

Esboite—an orbicular diorite in which andesine or oligoclase is the dominant plagioclase and forms the orbicles (Esbo, Finland).

Esterellite—a porphyritic quartz diorite with zoned andesine and hornblende (Esterel, France).

Kauaiite—an orthoclase-bearing olivine-augite diorite (or gabbro) with zoned feldspar (Kauai, Hawaii).

Kullaite—altered porphyritic diorite composed of phenocrysts of altered plagioclase and Na-orthoclase in an ophitic groundmass of feldspar and pseudomorphs of chlorite after augite (Kullen, Sweden).

Laugenite—an oligoclase diorite (Laugendal, Norway).

Luciite (*luclite*)—a plagioclase-hornblende diorite.

Mariupolite—an albite-nepheline-acmite-lepidomelane diorite (Mariupol, U.S.S.R.).

Ornöite—a hornblende diorite (Ornö, Sweden).

Ortlerite—a porphyritic hypabyssal diorite resembling greenstone (Ortles Alps, Italy).

Yukonite—a rock intermediate between a tonalite and an aplite (Yukon River, Alaska, U.S.A.).

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Cross-references: *Batholith*; *Calc-alkaline rocks*; *Intermediate igneous rocks*; *Plutonic rocks—classification and nomenclature*.

DOLERITE

Dolerite is the name given to the medium-grained intrusive basic igneous rock commonly found in dykes and sills; in North America and continental Europe it is often referred to as diabase, but many authors restrict this term to altered dolerite. Hatch et al. (1972) suggest that both of these terms be replaced by microgabbro (see **Basic igneous rocks**). Commercially, it is variously known as whin, whinstone, trap, or traprock (Carey, 1958; Walker, 1964). Ophitic and subophitic textures are common (Fig. 1). Quickly cooled varieties show porphyritic texture with phenocrysts of plagioclase, pyroxene and sometimes olivine.

Mineralogy and chemistry

Calcic plagioclase feldspar ($An_{>50}$) and augite are major constituents; iron ore (titanomagnetite) is found in subordinate amounts (Table 1). Some varieties are rich in olivine and these are referred to as olivine dolerites. Those with little or no olivine and containing silicic interstices (quartz-feldspar or glass) are referred to as tholeiitic dolerites (or quartz dolerites). Augite is commonly the only pyroxene present in the olivine dolerites, but in the tholeiitic dolerites it is generally accompanied by subcalcic pigeonite or hypersthene.

Normal undiversified dolerite is of very uniform chemical composition, closely resembling that of the equivalent basalt lavas, with the olivine dolerites and basalts having ca. 45%

TABLE 1. Modal analyses of dolerites

	1	2
Olivine	17.7	0.8
Pyroxene	19.8	32.3
Plagioclase	54.9	45.0
Iron ore	5.1	8.4
Interstitial residue	2.5 (zeolitic)	13.5 (silicic)

- 1. Olivine dolerites: average of 50 dykes from Islay-Jura swarm, W Scotland (Walker, 1960).
- 2. Tholeiitic dolerites: average of 26 intrusions in Central Scotland.

SiO_2 as against 50% in tholeiitic varieties (Table 2). Dolerites as a group show greater tendency towards diversification than basalts, but diversified types are omitted from Tables 1 and 2.

Diversification

Depending on the relative specific gravities of crystallizing minerals and magma, cumulate layers may be formed owing to minerals sinking toward the floors, or rising toward the roofs, of sills. Settling olivine may result in the formation of picrite layers near the bottoms of sills. Rising of early plagioclase crystals is less certain but, where assisted by rising convection currents, may explain some feldspar-rich layers at higher levels in sills. During crystallization, volatile-rich fluids may result in patches or veins of dolerite-pegmatite and, at a late-stage in the crystallization of tholeiitic dolerites, residual

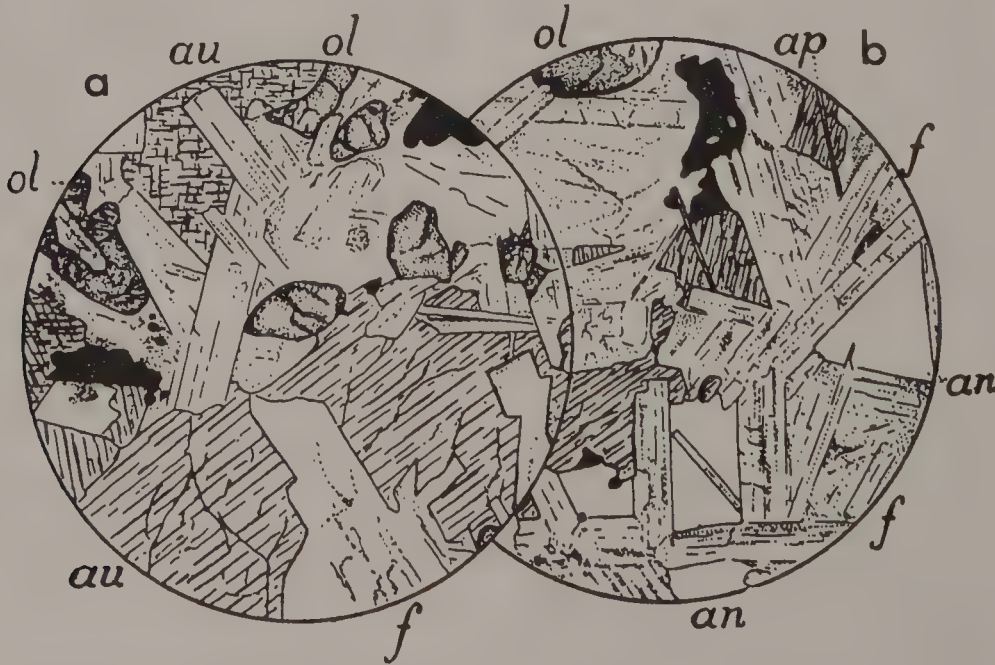


FIGURE 1. Microscopic texture of olivine dolerites; $\times 25$: (a) Antrim, Northern Ireland; (b) Arran, Scotland; an, analcite; ap, apatite; au, augite; f, feldspar (plagioclase); ol, olivine. (From Tilley et al., 1968.)

TABLE 2. Chemical data and norms for dolerites

	Chemical analyses		Trace elements		
	1	2		p.p.m.	
			3		4
SiO ₂	45.78	50.83	Rb	9	64
TiO ₂	2.63	2.03	Ce	21	123
Al ₂ O ₃	14.64	14.07	La	19	80
Fe ₂ O ₃	3.16	2.88	Sr	269	273
FeO	8.73	9.00	Ba	213	938
MnO	0.20	0.18	Ga	19	20
MgO	9.39	6.34	Nb	4	15
CaO	10.74	10.42	Cu	161	24
Na ₂ O	2.63	2.23	Zn	71	119
K ₂ O	0.95	0.82	Co	55	29
H ₂ O	0.76	0.91	Ni	96	9
P ₂ O ₅	0.30	0.23	Cr	163	0
	CIPW norms		Li	6	39
			V	253	130
			Sc	46	34
			Y	32	108
Quartz	—	3.5			
Feldspar	49.1	49.8			
Pyroxene	20.8	37.3			
Olivine	16.5	—			
Iron ore	9.6	8.0			
Others	3.3	0.5			

- 1. Average of 96 olivine dolerites and basalts (Nockolds, 1954).
- 2. Average of 137 tholeiites and tholeiitic basalts (Nockolds, 1954).
- 3. Average of 4 olivine dolerites, Ardnamurchan, W Scotland (Gribble, 1974).
- 4. Average of 4 tholeiitic dolerites, Ardnamurchan, W Scotland (Gribble, 1974).

fluids may produce small amounts of interstitial quartz-alkali feldspar intergrowths or glass. Variation of dolerite in dykes is rarer than in sills, but andesitic centers of dykes have been described (Edwards, 1942).

Dolerites may be slightly modified locally by the assimilation of country rocks such as shale. This is common in the Karroo province of S Africa, where most of the major sills are veined by mobilized (rheomorphic) sediment. Fusion of quartzofeldspathic sediments by dolerite may result in the formation of hybrids.

Other types

Allgovite (algovite)—a group of augite-plagioclase rocks ranging from dolerite through porphyritic varieties to gabbro (Allgäu Alps).

Beerbachite—a dyke rock, chiefly composed of labradorite, orthopyroxene, clinopyroxene and magnetite and with a granular (granoblastic) texture; it has been referred to as a metamorphosed dolerite, and as a hornfels, and occurs in some sheeted dyke complexes of ophiolites (Jelinek et al., 1980).

Calc-aphanite—a doleritic rock that has largely been replaced by carbonate minerals.

Cucalite—a chlorite-rich altered dolerite that grades locally into chlorite schist.

Devonite—a porphyritic dolerite containing large K-rich labradorite phenocrysts (Mt. Devon, Massachusetts, U.S.A.).

Granodolerite—a dolerite that contains quartz and orthoclase, usually as interstitial micropegmatite.

Hystorobase—a dolerite composed of plagioclase, quartz, biotite, and brown hornblende that is after augite.

Mimosite—a melanocratic dolerite containing abundant pyroxene and ilmenite.

Minverite—an altered dolerite containing primary hornblende and albite which may, in part, be primary (St. Minver, England).

Navite—a porphyritic variety of olivine dolerite containing phenocrysts of serpentinized olivine, labradorite, and augite in a groundmass composed chiefly of feldspar and augite.

Ophite—a general term for dolerites with ophitic texture, but uralitized pyroxenes, occurring in the Pyrenees.

Proterobase—a dolerite in which the mafic mineral is primary hornblende.

Soggendalite—a melanocratic dolerite containing abundant pyroxene.

Sordawalite—vitreous selvage of a dyke of olivine dolerite.

Uralitite—a dolerite in which the augite is altered to uraltite.

Wichtisite—modified glass-rich dolerite occurring as selvages in sills, dykes or other small intrusions (Wichtis, Finland).

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Cross-references: *Basalt*; *Basic igneous rocks*; *British Tertiary Volcanic Province*; *Dyke (dike)*; *Emplacement—modes*; *Sill*; *Textures of igneous rocks*.

DOUBLE-DIFFUSIVE CONVECTION

Double-diffusive convection is important in two geological environments: (1) on the ocean floor, where it prevents the dispersal of hot metalliferous volcanogenic brines, confining them to pools on the sea bed so that ore bodies can be produced, and (2) in magma chambers, where it plays a major role in the development of layering in layered intrusions.

It may occur when a body of liquid has both a compositional and a thermal gradient (Fig. 1). For example, if a layer of brine underlies a layer of fresh water, heating the base of the brine (A) will reduce its density and it will start to rise. However, the density will remain higher than

that of fresh water, so the brine will not be able to rise above the boundary with the overlying layer (B). A convection cell will be set up within the brine layer and heat will be passed upwards to the fresh water by conduction. This will cause the fresh water to convect also.

As convection causes mixing within each liquid layer, the boundary between the layers (L–L) becomes sharper. Diffusion of material across the boundary may eventually result in both liquid layers approaching the same composition, when they will combine to form a single layer. It is the diffusion of both heat and material across liquid–liquid boundaries that gives the name “double-diffusive” to this type of convection.

The thermal gradient needed for double-diffusive convection may result from cooling of the top of a magma chamber, and the compositional gradient can be produced by the replenishment of a fractionated magma chamber with fresh, unfractionated magma. In some cases where this has happened, diffusion may have destroyed the liquid layering before the magma chamber finally solidified. In the Skaergaard intrusion, however, the final consolidation occurred while the magma was still layered (McBirney and Noyes, 1979). Many of the rocks of the Layered Series in the center of this intrusion show large variations in composition (from basic to acidic) within a single layer, and so could not have formed by settling of crystals to the floor of the magma chamber. Their layering was originally formed in the liquid magma (Fig. 2). As the intrusion solidified from the base upward, crystallization of mafic minerals at the base of each layer left the top of the layer more acidic.

The liquid layering was preserved because the fractionation trend was tholeiitic. As a layer solidified, the residual liquid at the top of the layer became more Fe-rich and denser, increasing the density contrast with the overlying layer. In calc-alkaline intrusions, the upper part of each layer becomes less dense with fractionation and intrudes the mafic base of the layer above, so that the intrusion appears to be composite.

Double-diffusive fractional crystallization, with crystallization on the chamber floor releasing heat to drive convection, appears to have been a major factor in controlling the development of layered intrusions (McBirney and Noyes, 1979; Irvine, 1980; Irvine et al., 1983). While convection is occurring, crystallization can only take place in the basal liquid layer, since all higher layers are heated from below and remain superheated (Fig. 1). Any crystallization above the basal layer would result in the breakdown of the layering as the crystals

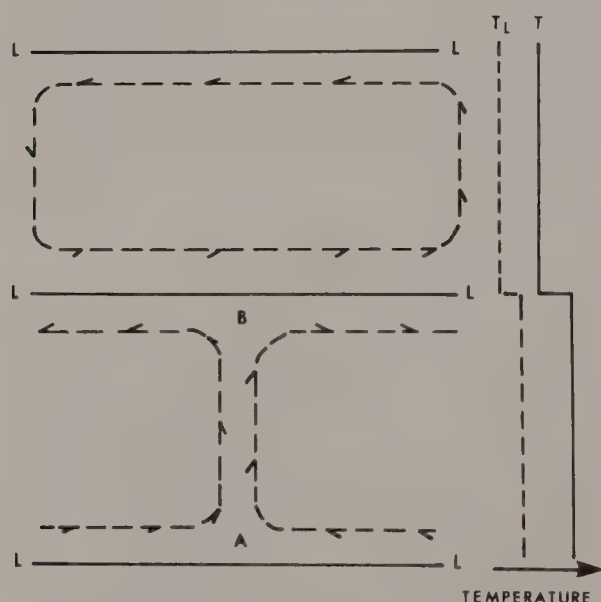


FIGURE 1. Double-diffusive convection in a layered liquid; layers are separated by diffusive interfaces (L–L) across which there is both a thermal and compositional gradient. The liquid heated at A can rise only to B before the less-dense material in the overlying layer arrests it. Throughout the liquid, the liquidus temperature (T_L) is lower than the actual temperature (T). Only in the basal liquid layer is it possible for these temperatures to be equal. Density distribution would be similar to the actual temperature distribution.

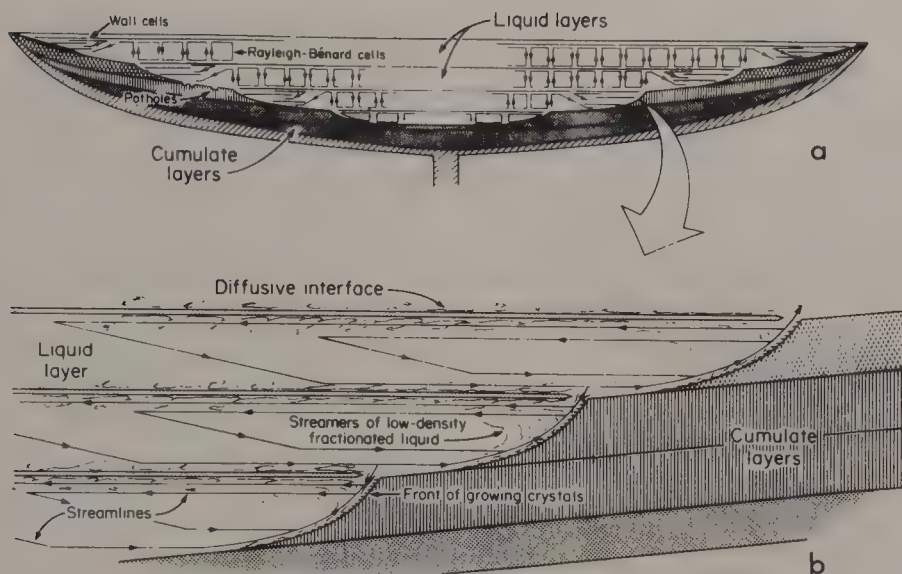


FIGURE 2. Crystallization in a layered intrusion: (a) solidification by lateral growth of cumulate layers in density-stratified magmatic liquids undergoing double-diffusive convection; (b) formation of a succession of cumulate layers where the liquids become less dense with crystallization. (From Irvine et al., 1983.)

settled. The more acidic the magma at the top of the chamber, the lower the liquidus temperature, and the more superheated it is likely to be. This may partly explain the greater abundance of xenoliths in acidic as compared with basic rocks.

I. W. FERGUSON

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Cross-references: *Bushveld Complex*; *Cumulate textures*; *Intrusion—layered*; *Liquid immiscibility*; *Magma*; *Magmatic currents and flow*; *Skaergaard intrusion*.

DURBACHITE

A biotite- and K-feldspar-rich mela-syenite was named a *durbachite* by Sauer (1893) after a

locality in the Schwarzwald (Black Forest) region of southern W Germany. Durbachites also occur in the Vosges Mtns. of eastern France and in the Moldanubian zone of the Bohemian Massif in both Czechoslovakia and Austria, where a large number of intrusions are concentrated in two linear NNE-trending zones (Fig. 1). Bodies of hypersthene–clinopyroxene–biotite mela-syenite with chemical compositions similar to those of durbachites also occur near these zones and the term *durbachitic rocks* is used to describe an apparently related suite of plutonic rocks that range from mela-granite and mela-syenite up to (rarely) ultramafic types. Most are amphibole–biotite mela-granites and (mela-)quartz syenites; less common are granodioritic types. High proportions of Mg-rich biotite and light-green amphibole are typical and the color index can be very high in rocks that approach amphibole biotites. Pseudomorphs after phenocrysts of pyroxene and olivine have been found in cognate xenoliths. Chemically the suite has affinities with K-rich lamprophyres and lamproites with high concentration of incompatible elements combined with high contents of elements typical of basic and ultrabasic rocks. All types of durbachitic rocks also display a prominent enrichment in light REE. These features distinguish the durbachite suite from the other Hercynian plutonic masses of central Europe (e.g., the gabbro–tonalite–granite series of the Central Bohemian pluton) and its origin is explained by derivation in the upper mantle of magma primarily enriched in incompatible elements (Holub, 1977, 1978). The composition of the

DYKE (DIKE)



FIGURE 1. Distribution of the main bodies of durbachitic rocks (dark) in the Moldanubian part of the Bohemian Massif, central Europe. (From Holub, 1977.)

more acidic members of the durbachite suite may be explained by the mechanism of magma mixing with crustal anatexitic melts.

F. V. HOLUB

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* Contains references to regional distribution and rock description.

Cross-references: *Appinite*; *Granite*; *Lamproite*; *Lamprophyre*; *Syenite*; *Ultramafic rocks*.

DYKE (DIKE)

A *dyke* (also spelled dike) is a tabular body usually of igneous rock discordant to, and

intrusive into, the bedding, schistosity, or other primary structure of its host rock (Fig. 1). Depending on the attitude of the host rocks, the dyke may be vertical, horizontal, or inclined, but most are vertical or steeply inclined, and were so intruded. Sediments, particularly clays and sands, may also be injected as dykes while in a plastic or “quick” state (Harms, 1965).

Igneous dykes occur in all sizes up to 200 km long and 100 m thick (Anderson, 1951) but most are only a few kilometers long and 1–5 m thick. The width of dykes is generally very constant over long distances, which indicates that regional crustal stresses participate in their formation.

Dyke walls are typically clean-cut and unsheared, and their displacement usually indicates only horizontal and tensional separation without faulting. Such dykes as are intruded along faults commonly have sheared and broken irregular walls and have variable thickness. These dykes also tend to carry inclusions of wall rock.

Dykes are emplaced as fluid or partially crystallized magmas that cool and solidify in periods of weeks or months. At depth, and especially near large igneous masses, cooling is slow and dyke rocks may be coarse-grained. Nearer the surface the rocks are medium- to



FIGURE 1. Basic dyke cutting thin bedded siltstone and showing an aureole of contact (thermal) metamorphism; near Melbourne, Australia.

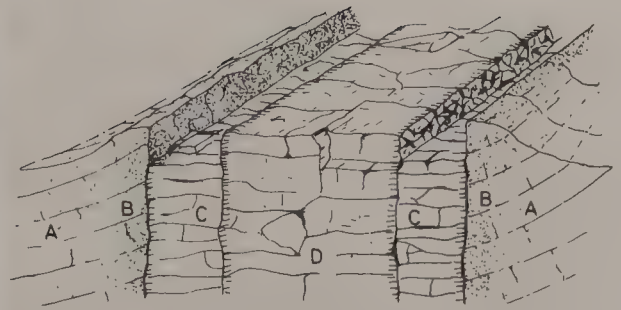


FIGURE 2. Composite dyke consisting of dyke D (often granophyre) intruded into dyke C (often dolerite) which is discordant to sedimentary bedding (A); a contact aureole (B) is shown.

composite dykes are of granophyre intruded into dolerite (Fig. 2).

Headed dyke—has a terminal expansion of tear-drop shape.

Pebble dyke—contains sedimentary fragments in an igneous matrix, the fragmentation having been caused by gaseous or aqueous fluids of magmatic derivation and rounding of the fragments caused by milling and/or corrosive action of hydrothermal fluids during injection upwards.

Secundine dyke—intruded into hot country rock.

Welded dyke—a secundine dyke whose boundaries have become obscured by mineral growth of country rock into the intrusion.

Dyke swarm—where large numbers of dykes were intruded at random, or in specific regions with prevailing stress systems that may develop a common or related pattern (Fig. 3).

fine-grained and acidic rocks commonly solidify as glass. Differential cooling may cause variation of grain size from coarse at the middle to fine at the margin of dykes. Movement of magma during crystallization may further cause the largest crystals to concentrate in the middle of the dyke and cause inclusions to be drawn into planar or linear fabrics.

Compositions of dyke rocks vary from ultrabasic to acidic, and common types are dolerite, lamprophyre, microgabbro, microdiorite, granophyre, aplite, and felsite. The coarsest-grained dyke rocks are pegmatites, which are transitional into vein deposits.

Dykes are injected as pulses of magma that may be so timed that each pulse solidifies as an individual dyke, or as successive pulses intruding the cores of dykes that have not yet completely crystallized. The following types and associations of dykes are recognized.

Simple dyke—compositionally uniform and generally represents a single pulse of magma (Fig. 1).

Multiple dyke—consists of several simple dykes of similar composition adjacent to and intruded alongside or within each other.

Composite dyke—consists of two or three units of different compositions intruded alongside or within each other; the most common



FIGURE 3. Distribution of early and middle Proterozoic dyke swarms and flood basalts in the Canadian Shield. (After Baragar, 1977.)

Radial dyke swarm—consists of dykes radiating from a volcanic center (Ode, 1957) that usually intrude only the volcanic pile and adjacent sediments, not the underlying basement rocks; they are caused by magma welling up tensional fractures produced by radial and outward pressure exerted by rising magma in a volcanic conduit.

Regional dyke swarms, which consist of numerous large and small dykes lying parallel to each other, represent magma (usually basic) that welled up fractures related to a tensional stress regime, often of transcontinental dimensions (Fig. 3); dolerite is the common rock type. Some are related to continental rifting and the early stages of continental break-up, e.g., the Tertiary coastal dyke swarms of E and W Greenland, related to the opening of the N Atlantic Ocean and Davis Strait, respectively (Myers, 1980). In the Precambrian, major transcontinental dyke swarms appear on all the cratons in the Proterozoic. They are considered to reflect a stage in craton growth and maturity when the cratonic nuclei became stable enough to transmit regional tensional stress. Examples of these swarms, which occasionally show a relationship to flood basalt provinces, include the Mackenzie swarm in Canada, the Waterberg dykes of southern Africa, and the Jatulian dykes of the eastern Baltic Shield (see Windley, 1984; Halls and Fahrig, 1988).

Giant dykes occur in W Greenland where they form part of a regional swarm related to the middle Proterozoic Gardar igneous province and rift zone (Bridgewater and Coe, 1969). The individual dykes are over 100 m wide and, though vertical structures, they often display subhorizontal layering. Some show evidence of multiple intrusion. Similar giant dykes are known from W Australia (e.g., Jimberlana dyke)—and southern Africa (Great Dyke of Zimbabwe). Some may be related to stratiform layered basic intrusions, some of which have dyke-like lower feeder zones (e.g., Great Dyke, Zimbabwe; Muskox intrusion, NW Canada).

Sheeted dyke complexes form a part of most ophiolites (with important exceptions—Coleman, 1977) and are considered to represent disrupted fragments of layer 2 of former oceanic crust. A sheeted dyke complex consists largely or wholly of dykes, usually dolerite (commonly altered) with minor sheets of gabbro, beerbachite, microdiorite, trondhjemite, or granophyre.

Where screens of wall-rock remain, this is either basalt or spilite lava (often pillow lava), or gabbro, depending on whether it is the top or base of the complex that is examined. At the top of a sheeted dyke complex, dykes can be traced into feeder tubes for pillow lavas or small sills. These complexes form beneath the rift zone on a mid-oceanic or back-arc basin spreading axis (Cann, 1974; Coleman, 1977; Pallister, 1981). Like the overlying lavas, the sheeted dyke complex usually shows the effects of ocean-floor metamorphism.

JOHN BRADLEY

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Cross-references: *British Tertiary Volcanic Province*; *Dolerite*; *Emplacement—modes*; *Granites and mineralization*; *Intrusion*; *Magma*; *Ophiolite*; *Pegmatite*; *Ring dyke*; *Volcanoes and volcanism*.

E

EARTH'S CRUST GEOCHEMISTRY— See Vol. IVA

ECLOGITE

The rock name *eclogite* was first used by R. J. Haüy in 1822 for a “fancy rock” composed mainly of two minerals “diabase and garnet accompanied accidentally by kyanite, quartz, epidote, and lamellar amphibole.” Later the name was applied to all granular rocks consisting essentially of greenish clinopyroxene and reddish garnet, with a possible admixture of several primary and secondary minerals. The rock is conspicuous for its high density (ca. 3.5 g/cm³), compared to basalt with approximately the same chemical composition (ca. 3.0).

Eclogites are rather rare rocks occurring as relatively small intercalations, lenses, boudins, or enclosures within various hosts. They are most common in gneissic or migmatite com-

plexes in close association with amphibolites, and they were considered by F. Becke and U. Grubenmann to be basic bodies catatonally transformed under very high pressures, which resulted in the rocks assuming the smallest possible specific volume. Rocks of similar appearance are also found within peridotites or as nodules in the kimberlite pipes of South Africa. Eskola (1921) therefore considered them to be the crystallization products of basic magmas in very deep zones, under very high pressures and temperatures, to which he assigned a separate eclogite mineral facies.

Smulikowski (1964) and Coleman et al. (1965) considered that eclogites could be classified according to their mineral and whole-rock chemistry, and that this also corresponded to their geological setting. Smulikowski (1968; 1972), using more recently published chemical analyses, further refined this classification (Fig. 1; Table 1).

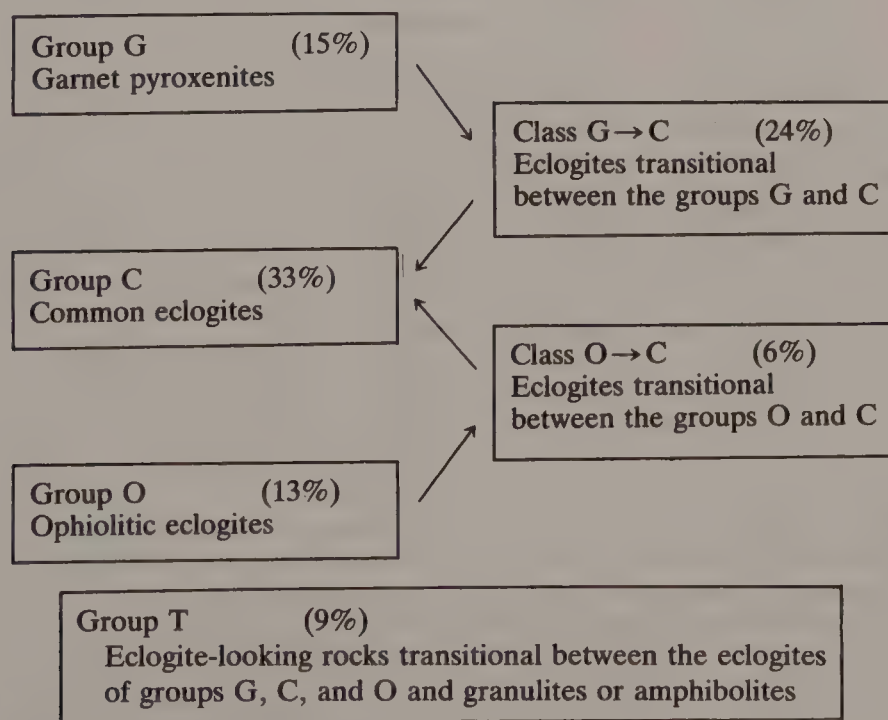


FIGURE 1. Classification scheme and various proportions (based on 287 analyses) of eclogite and eclogite-looking rocks showing three primordial groups (G, C, O), two classes transitional between the groups G→C and O→C, and a separate Group T of plagioclase-bearing eclogite-like rocks constituting transitions between the groups G, C, and O and granulites or amphibolites.

TABLE 1. Average chemical compositions of rocks, garnets and clinopyroxenes of the eclogite groups distinguished in Fig. 1

	Group G	Transit. G-C	Group C	Transit. O-C	Group O	Group T
Rocks						
SiO ₂	47.2	45.4	48.2	47.2	47.3	47.8
TiO ₂	0.5	1.5	1.0	2.7	2.5	1.2
Al ₂ O ₃	11.4	15.3	16.8	13.4	13.0	16.5
Fe ₂ O ₃	2.9	3.2	2.4	3.4	5.5	3.3
Cr ₂ O ₃	0.3	—	—	—	—	—
FeO	6.7	7.6	7.8	9.3	8.8	7.5
MnO	0.3	0.2	0.1	0.2	0.2	0.2
MgO	16.4	11.7	8.7	8.2	5.9	9.0
CaO	12.6	12.6	11.5	10.9	10.6	10.5
Na ₂ O	0.7	1.4	2.3	3.0	4.0	2.1
K ₂ O	0.2	0.3	0.4	0.4	0.4	0.5
P ₂ O ₅	0.1	0.1	0.1	0.7	0.3	0.2
H ₂ O	0.9	0.9	0.7	0.9	1.5	1.2
TOTAL	100.2	100.2	100.0	100.3	100.0	100.0
Garnets						
SiO ₂	41.4	40.2	39.3	39.2	37.7	39.0
Al ₂ O ₃	22.3	21.5	21.5	21.3	20.1	21.1
Fe ₂ O ₃	1.6	2.1	2.5	1.8	2.4	2.3
FeO	10.8	18.7	18.6	20.9	24.9	20.3
MgO	17.2	11.0	9.1	7.1	3.4	7.3
CaO	5.5	6.0	8.5	8.9	9.9	9.1
MnO	0.4	0.4	0.3	0.6	0.9	0.5
Cr ₂ O ₃	0.8	0.1	—	—	—	—
TiO ₂	0.2	0.4	0.4	0.1	0.7	0.4
H ₂ O	0.2	trace	0.3	0.2	trace	0.2
TOTAL	100.4	100.4	100.5	100.1	100.0	100.2
Clinopyroxenes						
SiO ₂	52.9	53.2	53.8	53.1	53.9	50.6
Al ₂ O ₃	3.6	6.6	10.0	8.7	9.5	9.1
Fe ₂ O ₃	1.2	2.6	1.8	4.6	6.1	2.3
FeO	2.5	5.0	2.3	3.2	3.2	5.2
MgO	16.2	11.2	10.4	9.2	7.1	9.8
CaO	21.5	16.9	16.5	15.1	12.7	19.1
Na ₂ O	1.1	3.8	4.2	4.8	6.5	2.5
K ₂ O	0.1	0.1	0.2	0.1	0.1	0.1
MnO	0.1	0.1	trace	trace	0.2	0.1
Cr ₂ O ₃	0.4	—	—	—	—	—
TiO ₂	0.2	0.4	0.4	0.3	0.4	0.8
H ₂ O	0.6	0.4	0.3	0.7	0.3	0.5
TOTAL	100.4	100.3	99.9	99.8	100.0	100.1

Classification

Group G—garnet pyroxenites. These are characterized by strongly femic chemistry. Their garnets display the highest proportion of MgO, and the lowest proportion of Fe. Clinopyroxenes are very poor in Na, with the normative jadeite molecule in the whole rock remaining below 15 mol% (Fig. 2). Thus, the clinopyroxenes do not correspond to omphacite, typical in true eclogites. Orthopyroxene, olivine, pale brownish primary hornblende, and sometimes biotite or spinel frequently occur as accessory

minerals.

Varieties include *griquaïtes* (orthopyroxene-free) and *garnet websterites* (orthopyroxene-rich), with Cr-bearing diopside or endiopside. They are found frequently in the garnet peridotites or garnet serpentinites of the Bohemian Massif (Austria and Czechoslovakia), the peridotites of SW Norway, and in the kimberlite pipes of S Africa (including *newlandite*, a griquaïte containing garnet, enstatite, and chrome diopside). Garnet ariégites accompany Pyrenean lherzolites, and contain Na-poor augite and some spinel.

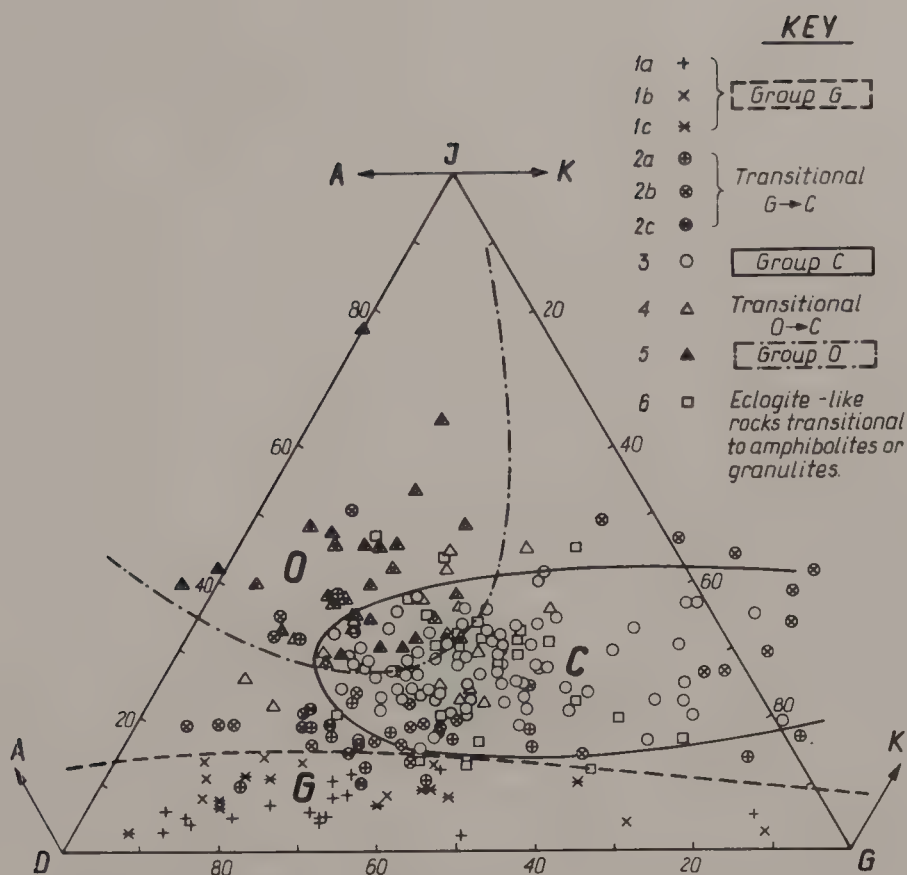


FIGURE 2. Eclogite mineral norm J-D-G (A or K) with coordinates of the molecular proportions of normative minerals calculated from the bulk chemical analyses of rocks: J, jadeite ($\text{NaAlSi}_2\text{O}_6$); D, diopside ($\text{Ca}(\text{Mg}, \text{Fe})\text{Si}_2\text{O}_6$); G, garnet ($\text{Fe}^{2+}, \text{Mg}, \text{Ca}$) $_3\text{Al}_2\text{Si}_3\text{O}_{12}$. In special cases it may become necessary to introduce additional minerals: for rocks deficient in Al_2O_3 ; A, acmite ($\text{NaFe}^{3+}\text{Si}_2\text{O}_6$), and for rocks with Al_2O_3 excess; K, kyanite (Al_2SiO_5).

Garnet pyroxenites connected with 1a (+), peridotites; 1b (x), kimberlites; and 1c (x), alkali-basaltoid volcanic rocks. Eclogites transitional between groups G and C connected with 2a (⊕), peridotites; 2b (⊗), kimberlites; and 2c (⊗), alkali-basaltoid volcanic rocks. 3 (○), Common eclogites. 4 (△), Eclogites transitional between ophiolitic and common eclogites. 5 (▲), Ophiolitic eclogites. 6 (□), T group of plagioclase-bearing eclogite-like rocks.

G group rocks (broken line) plot in the field along the DG side of the triangle, below 15% jadeite molecule; C group common eclogites (solid line) occupy the central part of the triangle extending to the right and transgressing its JG side; O group (chain line) of ophiolitic eclogites concentrates in the upper left part of the triangle tending to overstep the JD side, but, on the other hand, overlapping to some extent the field of the C group; transitional G→C eclogites plot chiefly at the peripheries of the common eclogite field, in the garnet-poor sector on the left side and in the kyanite sector on the right side as well.

Garnet-augite or garnet-fassaite pyroxenites, with or without orthopyroxene, olivine, brownish hornblende, are found as nodules in alkali basalts and nephelinite lavas and pyroclastic rocks on Oahu (Hawaiian Islands), Delegate (Australia), and Kakanui (New Zealand).

According to Banno and Matsui (1965), very low partition coefficients for Mg-Fe between garnet and clinopyroxene in these rocks indicate very high temperatures for crystallization (average $K_D = 4.1$). These rocks are considered to originate in the upper mantle.

Group C—common eclogites. These have a basaltic chemical composition. Their garnets display a wide variability in their Fe^{2+}/Mg ratio, but the prevalence of Mg over Fe^{2+} , or Fe^{2+} over Mg is never very great (Fig. 3). Clinopyroxene is always represented by omphacite with a moderate amount of the jadeite molecule; the Al content usually exceeds that of Na (Fig. 4). Pale brown, or brown-olive green primary hornblende (carinthine), zoisite, quartz, kyanite, white mica (Mg-phengite), rutile, and sometimes primary dolomite all

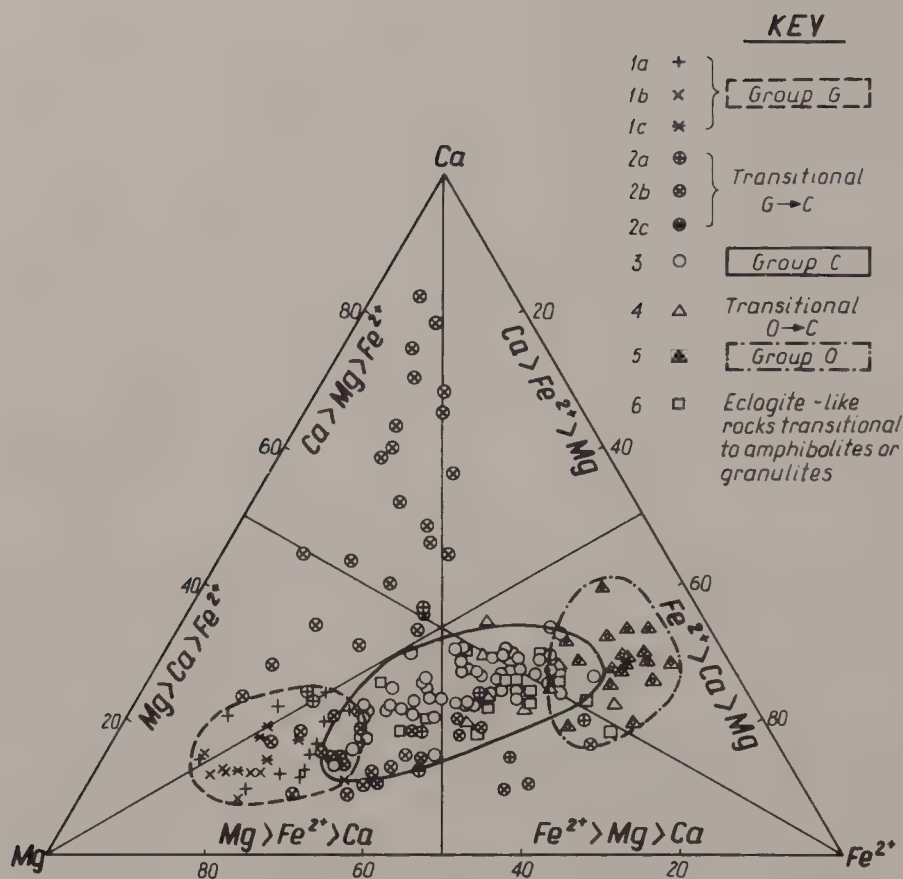


FIGURE 3. Garnet composition on Mg-Fe²⁺-Ca plot; key as in Fig. 2; plots of individual garnets from rocks belonging to different rock groups concentrate in different fields which to some extent overlap.

The garnets of Group G plot near the Mg-corner. Garnets of Group O are concentrated in a field approaching the Ca-Fe side line. Garnets of common eclogites (Group C) are scattered in a long elliptical field intermediate between the fields of the G and O groups. Garnets of transitional G→C eclogites plot at the peripheries of the main field of common eclogite garnets at rather low Ca and Fe contents; towards the Ca corner are the E Siberian kyanite eclogite-grospydite series (Sobolev et al., 1968). Garnets of the O→C transition class and of the T group are scattered chiefly in the Fe-rich half of the C group garnet field, often approaching the O group field.

occur as accessory phases. The Fe-Mg partition coefficients between garnet and omphacite vary within broad limits, but an average coefficient of $K_D = 9.4$ points to moderate temperatures of crystallization.

Eclogites of Group C are most common as intercalations, lenses or boudins in gneissic or migmatite complexes of supracrustal origin, metamorphosed in the amphibolite or granulite facies. They are commonly wholly or partly amphibolitized and behave like metastable relics from an earlier stage of high-pressure metamorphism. They may originally have formed dolerite dykes or sills, or basalt lava sheets within sandstone and mudstone sequences, later transformed to gneisses or migmatites. Some eclogite intercalations may be of sedimentary origin, formed from dolomite-rich mudstone mixed with basaltic pyroclastic deposits.

The best-known occurrences of this type of eclogite are those of Fichtelgebirge, W Ger-

many; Saxony, E Germany; the Polish Sudetes; Saualpe, Austria; lower Loire district, France; Inverness-shire, Scotland; western Norway; and the northern Ural mountains, U.S.S.R.

Group O—ophiolitic eclogites (Alpine-type). These display more Na₂O, Fe₂O₃, FeO, but less Al₂O₃ and MgO in their bulk composition than groups G and C. Their garnets are particularly poor in Mg, their clinopyroxenes are Na-rich omphacites or chloromelanites with Na equal to or exceeding Al (Fig. 4). Their bulk chemical composition corresponds to strongly alkaline basalts or spilites. The most characteristic supplementary minerals are glaucophane, clinozoisite, or epidote, and white mica (phengite or paragonite). Ti minerals, chiefly rutile, but also ilmenite and sphene, are more abundant than in groups G and C.

The eclogites of this group occur as intercalations or lenses in blue schists and tectonic mélange; accompanying rocks are

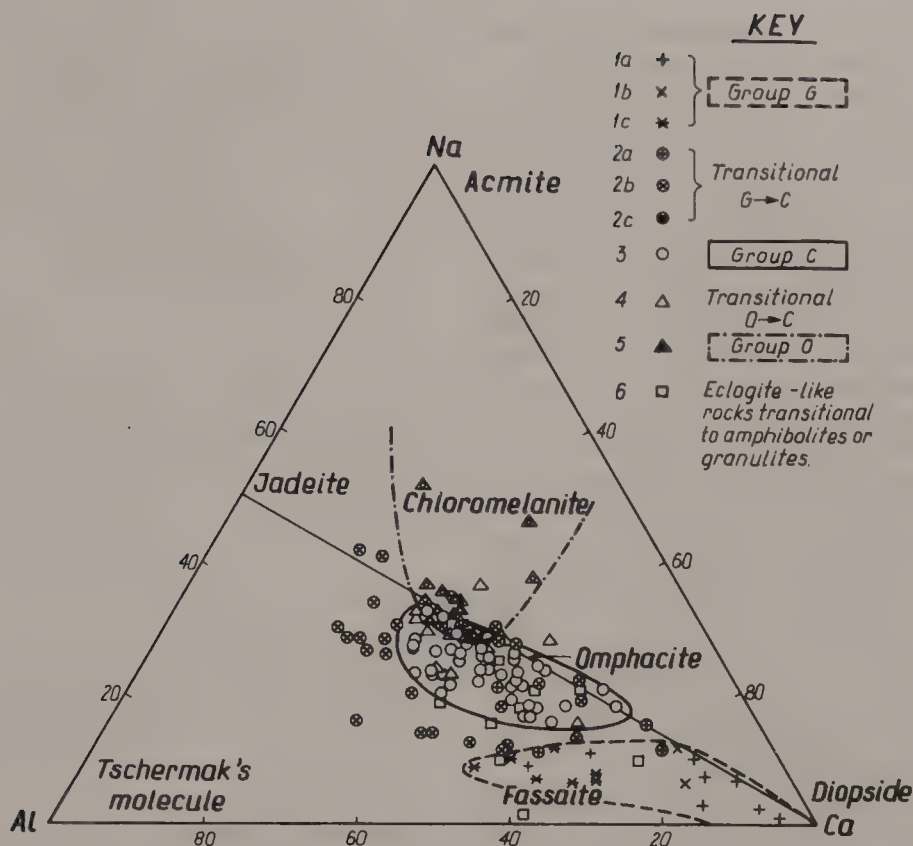


FIGURE 4. Plot of clinopyroxene compositions: key as in Fig. 2. Clinopyroxenes of G group are always low in Na; in the rocks connected with peridotites or kimberlites they usually approach endiopsidite, while in nodules of alkali-basaltoid volcanic rocks they are more or less Al-rich and approach augite or fassaite. Common eclogite omphacites concentrate in a well-defined field displaced distinctly from the jadeite–diopside line toward the Al corner. The strongest displacement in this direction is proper to grosspyrite omphacites, while omphacites of transitional G→C eclogites take usually low Na positions. Omphacites of ophiolitic eclogites contain equal proportions of Al and Na, or they display some excess of Na, signifying some acmite participation and thus chloromelanite chemistry. The field of O group clinopyroxenes occupies the highest position in the triangle, although it overlaps, to a small extent, the field of common eclogites. Omphacite plots for the T group of rocks are, as a rule, situated lower than those of common eclogites, because Na must here be subject to partition between clinopyroxene and plagioclase.

serpentinites and such metasediments as calcareous slates, metagraywackes, and meta-cherts. This rock assemblage is often associated with ophiolite obduction and the subduction of oceanic crust and reflects low-temperature–high-pressure metamorphism. High Fe–Mg partition coefficients between garnet and clinopyroxene (average $K_D = 15.8$) supports such conditions of crystallization.

Well-known sites for this type of eclogite are in the Franciscan rocks of California, U.S.A.; El Rosario, Guatemala; Guajira Peninsula, Colombia; several localities in Japan; the Sesia-Lanzo (Ivrea Zone) area of Piedmont, Italy; the Wallis Alps, Switzerland; and Shubino, southern Urals, U.S.S.R.

Transitional eclogites between groups O and C. These have been found in many localities in the eastern Alps (Austria), Piedmont (Italy), S

Brittany and the lower Loire region (France, Massif Central), in some parts of W Norway, and Shikoku, Japan. The rocks are similar to the O group in their association with glaucophane-bearing blue schists, and in their Na-amphibole and epidote content, but their garnets and omphacites are not typical of the O group and approach those of Group C. The chemical composition of the protolith and probably also the P – T conditions of recrystallization were intermediate between those of groups O and C.

Transitional eclogites between groups G and C. These are more diversified, because each of the three occurrences of these garnet pyroxenites (those connected with peridotites, forming nodules in kimberlites, and xenoliths in nephelinites) represent a different transition to common eclogites. Transitional eclogites genet-

ically connected with peridotites are particularly frequent in the Moldanubian granulite terrane (Bohemian Massif) of Czechoslovakia and Austria. Their garnets approach the composition of those in Group C, but their omphacites are usually Na-poor, frequently approximating to augite; Al being low precludes the appearance of kyanite. In contrast, the eclogite nodules from S African kimberlites contain usually a very magnesian garnet, but a rather normal Na-rich omphacite; their Al content may be either low or very high giving rise to abundant kyanite.

One of the E Siberian kimberlite pipes contains a peculiar eclogite variety, *grosphydite*, composed of a grossular-rich garnet, abundant kyanite and omphacite of normal Na content, but with excessive Al. It was suggested that grosphydites originated from anorthosites under very high pressure. It appears quite possible that the whole transition group originated from different crustal basites that have been thrust down into the upper mantle and recrystallized there under very high pressures.

Group T—eclogite-like rocks transitional between groups G, C, and O and other metamorphic rocks. These rocks are characterized by the presence of some plagioclase in primary paragenesis with garnet and clinopyroxene. The latter may be of normal omphacite composition, but more frequently it is rather poor in Na and rich in Al^{IV} , more or less approaching augite (Fig. 4). The chemical composition of the whole rocks and the included garnets approach those of common eclogites. The rocks may belong to the granulite facies and present a transition from the high-pressure clinopyroxene–garnet granulites to eclogites. Typical examples have been described from the Polish Sudetes and from Cabo Ortegal in NW Spain. Some other representatives belong to the amphibolite facies; they contain primary oligoclase and hornblende, sometimes zoisite, kyanite, and phengite in equilibrium with garnet and omphacite, and they represent a transition between amphibolites and hornblende eclogites. The best-known examples of this type occur in the Polish Sudetes.

Genesis

The origin of eclogitic rocks may be interpreted in accordance with the experiments of Green and Ringwood (1967), demonstrating that the gabbro–eclogite transformation proceeds gradually with a rise of pressure, producing several intermediate parageneses of clinopyroxene–garnet granulites. The plagioclase-bearing rocks of Group T may represent products of incomplete “eclogitization” of primitive basites

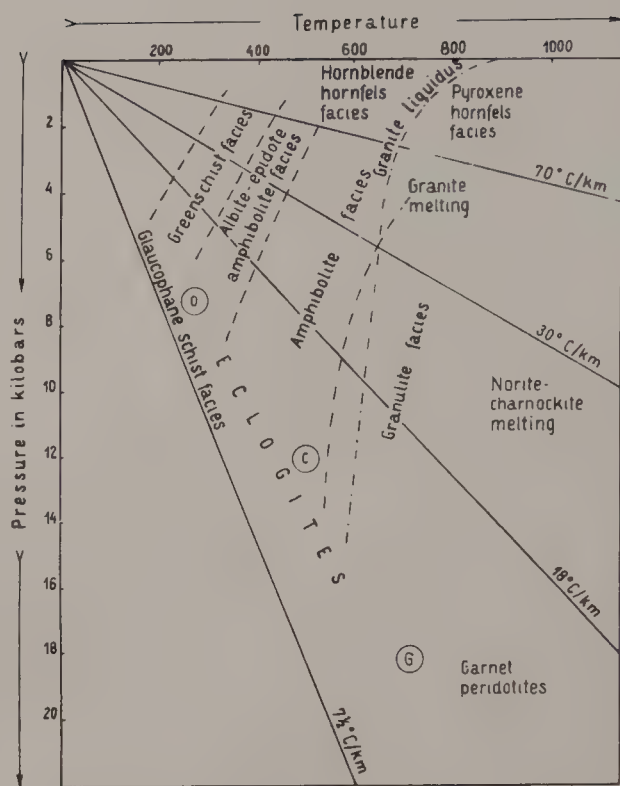


FIGURE 5. Eclogites in relation to different facies of regional metamorphism.

as a consequence of insufficient pressures, such as those at higher temperatures and low P_{H_2O} in the granulite mineral facies, and at somewhat lower temperatures and at higher P_{H_2O} in the amphibolite mineral facies.

In the P – T diagram of different mineral facies (Fig. 5) eclogites occupy a long belt in the very high pressure sector, ranging from 5 kb at 250°C up to 20 kb at about 800°C. Thus, not all eclogites represent the product of recrystallization in the eclogite metamorphic facies.

K. SMULIKOWSKI

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Cross-references: *Amphibolite*; *Amphibolite facies*; *Basic igneous rocks*; *Eclogite facies*; *Granulite*; *Granulite facies*; *Kimberlite*; *Nephelinite*; *Ophiolite*; *Peridotite*; *Pyroxenite*; *Upper mantle—experimental petrology*; *Xenolith*.

ECLOGITE FACIES

The *eclogite facies* is based on the distinctive association of omphacite pyroxene and pyrope-rich garnet forming the rock-type eclogite. Knowledge of rock types other than basic rocks in this facies is very sparse. Yoder and Tilley (1962) give possible assemblages (Fig. 1). Omphacite–garnet, omphacite–garnet–hypersthene, and omphacite–garnet–kyanite are common parageneses. In those eclogites associated with glaucophane schists, glaucophane is common. Feldspar is normally very minor or absent, but it would appear that anorthite is stable from certain original compositions, such as anorthosite, which is found in places associated with eclogites with the minerals corundum, sapphirine, and zoisite frequently associated with the anorthite. Rare plagioclase eclogites are found. In certain eclogites, other minerals such as phengite, lawsonite, scapolite, brown hornblende, and albite may be found, but these are confined to

certain associations. Rutile is an almost constant accessory and both apatite and pyrite appear to be stable universally. Other rock types probably of comparable facies are garnet peridotites, some peridotites, and ariégites.

Eclogites occur in several different geological environments and these have different minor mineral assemblages and the compositions of the garnet and pyroxene differ. They have been divided on this basis into three main groups by Coleman et al. (1965).

Group A: as nodules in kimberlite; as nodules in basic lavas; and as lenses in ultramafic bodies.

Group B: in granulites; in amphibolite facies gneiss and migmatites; in green-schists.

Group C: in glaucophane schist; in serpentinite.

Mineralogy

Group A. Pyrope content of garnet >55%; jadeite content of pyroxene >50%; a high relative proportion of the hypersthene molecule in the clinopyroxene; minor constituents are kyanite, lawsonite, plagioclase (An_{18-24}), scapolite, phengite, and diamond.

Group B. Pyrope content of garnet 30–55%; omphacites have >55% jadeite; minor constituent minerals differ according to the facies of the rocks in which they are found—brown hornblende and biotite in granulite facies; plagioclase, brown hornblende, and biotite in amphibolite facies; and actinolite, glaucophane, and paragonite in greenschist facies.

Group C. Pyrope content of garnet <30%; omphacites have >50% jadeite; minor constituents are glaucophane, lawsonite, and phengite.

Conditions of formation

The very high packing index (Fairbairn, 1943) and packing % (Poldervaart, 1953), which are the highest recorded for common rock types, have long been considered as evidence that eclogites are high-pressure assemblages. Experimental work has confirmed this (Kennedy, 1959; Yoder and Tilley, 1962), but other work (see Wyllie, 1971) shows that the boundary of the stability of eclogite is far from certain and that slightly different compositions (i.e., tholeiitic as distinct from alkali basaltic) apparently have widely different stability fields (Fig. 2).

At high temperatures, the eclogite field is bounded by the basalt melting curve and by a small field (Fig. 2) in which garnet is not stable and pyroxene-rich basic rocks are formed—the *pyroxenite facies* of Yoder and Tilley (1962) and

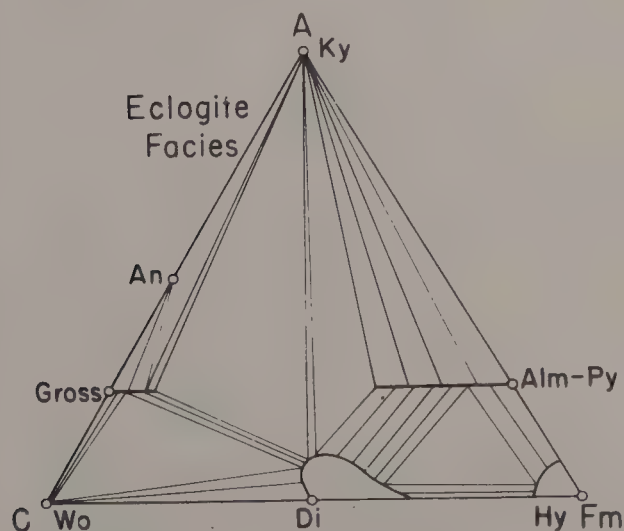


FIGURE 1. ACF diagram of eclogite facies. (After Yoder and Tilley, 1962.)

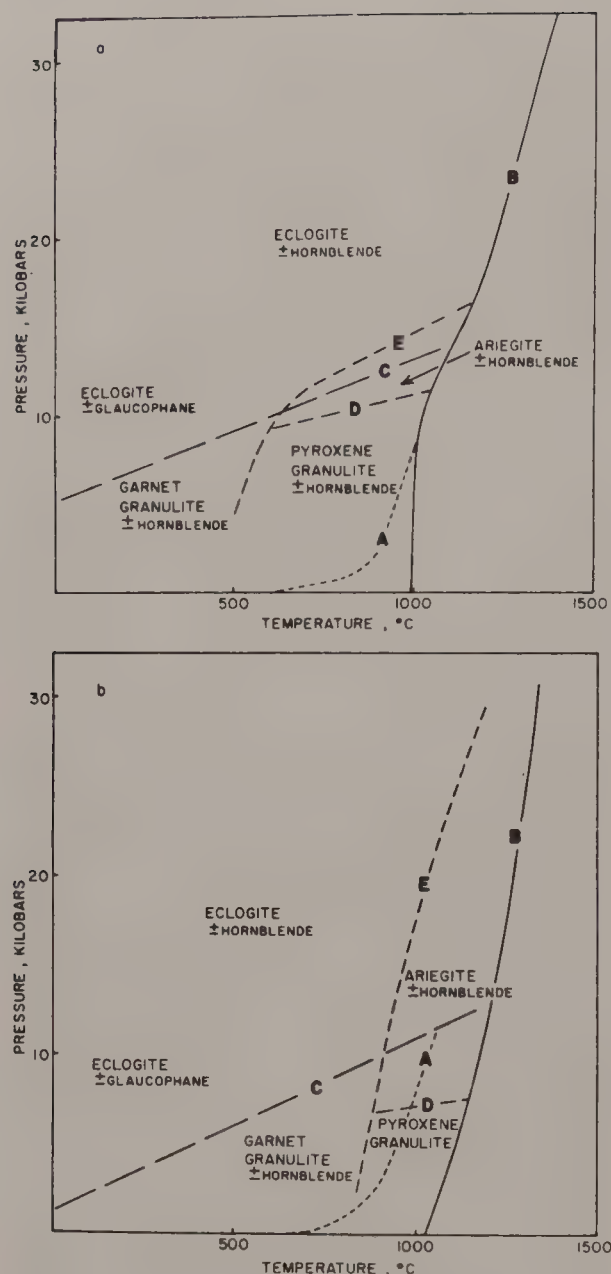


FIGURE 2. P - T diagrams indicating the stability field of eclogite formed from (a) silica-saturated and (b) silica-undersaturated basaltic rocks: boundaries of mineral stabilities are A, amphibole; B, basalt melting; C, omphacite; D, plagioclase; E, garnet. (From Church, 1968.)

ariégite facies of Church (1968). The stability of garnet varies with different rock compositions and the resulting pyroxene in ariégite also differs. From tholeiitic basalts it is aluminous clinopyroxene, and from alkali basalt it is clinopyroxene and spinel. Orthopyroxene, magnetite, and plagioclase are also stable.

The boundary of the eclogite facies, which is largely controlled by pressure, has a positive slope to the temperature axis so that at low temperatures eclogite forms at relatively lower pressures. Yoder and Tilley (1962) suggest figures of ca. 7 kb for temperatures of

200–300°C instead of about 10 kb at 1,000°C. Eclogite is thus stable down to quite low temperatures and is the high-pressure boundary of all the normal metamorphic facies, i.e., glaucophane schist, greenschist, amphibolite, and granulite facies, and is found as nodules and layers in all these major facies. The three groups of Coleman et al. (1965) can thus be regarded as different subfacies largely differing in temperature. Group A, which is not associated with normal metamorphic rocks, probably forms in the mantle at very high temperatures and pressures.

The eclogite mineral facies is defined in basic rocks, but a partial equivalent has been defined in metasediments, and tentatively named “white schist” (Fig. 3; Schreyer, 1973). The distinctive mineral parageneses have only been defined in Mg–Al-rich schists, and theoretical and experimental parameters are based on the systems $MgO-Al_2O_3-SiO_2-H_2O$ and $K_2O-MgO-Al_2O_3-SiO_2-H_2O$ (Schreyer and Seifert, 1969). “White schist” is effectively a Mg–Al-rich rock in which Mg–Al silicates (cordierite, gedrite, and Mg-chlorite) are replaced by coexisting Mg- and Al-silicates, usually talc + kyanite or talc + sillimanite. The transition at low temperatures (<600°C) appears to be directly from Mg-chlorite to talc + kyanite, but at high temperatures the breakdown of Mg-cordierite involves the production of intermediate silica-poor phases such as yoderite ($Mg_4Al_{10}Si_7O_{30}(OH)_4$) and sapphirine ($Mg_2Al_4SiO_{10}$). Accessory phases typical of these rocks include phlogopite,

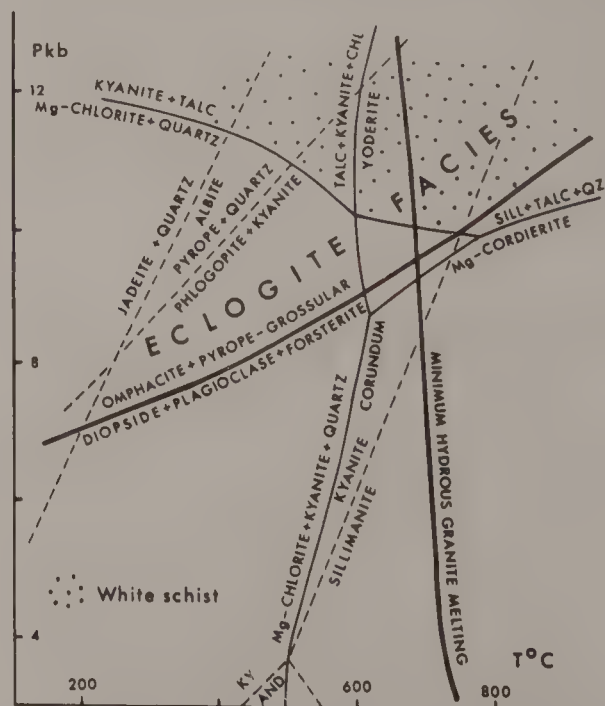


FIGURE 3. P - T diagram for mineral reactions defining the “white schist” facies. (After Schreyer and Seifert, 1969.)

pyrope garnet, corundum, kornerupine, zoisite, and spinel. Dehydration involves the production of enstatite and olivine.

Rocks in this mineral "facies" have been described from Greenland (Fiskanaeset), S Africa–Zimbabwe (Messina, in the Limpopo belt) and Afghanistan (Sar-e-Sang) (see Schreyer and Abraham, 1975, 1976; Friend, 1982).

A. E. WRIGHT

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Cross-references: *Amphibolite facies*; *Eclogite*; *Glaucophane schist facies*; *Granulite facies*; *Mafic and ultramafic inclusions in igneous rocks*; *Metamorphic facies series*; *Peridotite*; *Pyroxenite*; *Upper mantle—experimental petrology*.

ELEMENT PARTITIONING IN PETROGENESIS

Element partition coefficients can be used to assist in solving problems related to major- and trace-element distribution, diffusion, and concentration during the processes of anatexis, fractional crystallization, and metamorphic

recrystallization (e.g., McIntire, 1963; Allègre and Minster, 1978).

The practical use of partition (distribution) coefficients in igneous and metamorphic petrogenesis is governed by the overall chemistry and mineralogy of the rock under consideration, and whether the element distribution being investigated is major or trace in terms of its concentration. If the elements under consideration are major lattice components in the mineralogical make up of the rock (whose substitution will be governed by Raoult's law), then their distribution may be modeled in the form of a solid solution between two phases in equilibrium, given that there is a possible element substitution. If the phases are in equilibrium, then the chemical activity of an element is equal in both phases (Berthelot–Nernst law). If an element is present in trace concentrations (perhaps up to 10,000 p.p.m.), then its substitution is governed by Henry's law. In this case the mole fraction of the element is still directly proportional to its chemical potential, but there is an additional composition-dependent term in the solution for chemical potential that is modified by the Henry's law constant.

Partition coefficients have their roots firmly within thermodynamic theory: with reference to the Berthelot–Nernst law, the chemical activity of a crystallizing end member in an equilibrium assemblage is related to its chemical potential (and therefore the Gibbs energy of the phase) by

$$\mu_{1A} = \mu_{1A}^{\circ} + RT \ln a_{1A}.$$

μ_{1A}° is the standard chemical potential of end member 1 in phase A; a_{1A} is the activity of end member 1 in phase A (the activity a_{1A} is a function of X_{1A} —the mole fraction of end member 1 in phase A). In the Henry's law region of substitution (at low concentrations or where $X_{1A} > 0$) a plot of $\ln X_{1A}$ vs. μ_{1A} gives a straight line (slope RT) with the equation

$$\mu_{1A} = G_{1A} + RT \ln X_{1A} + RT \ln h_{1A(2)}$$

(Fig. 1; Powell, 1978, p. 46) where $h_{1A(2)}$ is the Henry's law constant for end member 1 in phase A infinitely diluted by end member 2. In the Raoult's law region where $X_{1A} = 1$, to moderate dilution with end member 2, the relationship between μ_{1A} and $\ln X_{1A}$ is still linear (slope RT) and is given by

$$\mu_{1A} = G_{1A} + RT \ln X_{1A}.$$

Experimental determination of trace element substitution in the Henry's law region demonstrates that in practice there is a deviation from the ideal linear relationship that may be the result of defects within crystal lattices at the

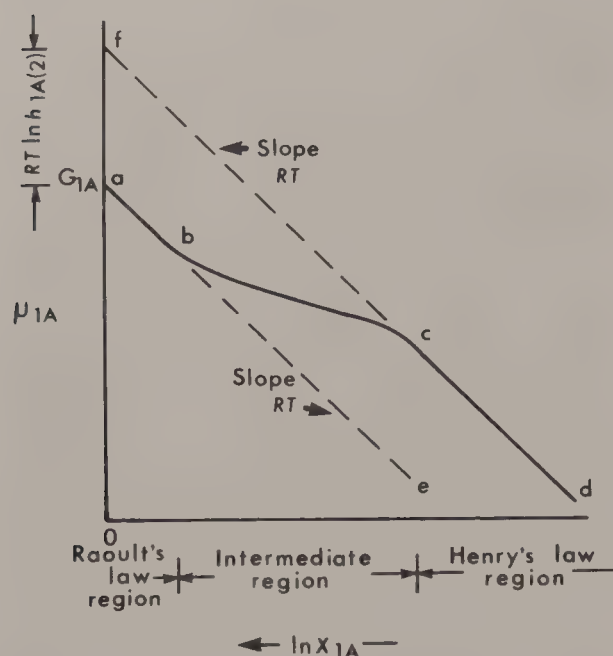


FIGURE 1. A plot of μ_{1A} against $\ln X_{1A}$ showing the parallel linear relationship between chemical potential and composition in Raoult's law region (intercept G_{1A}) and Henry's law region (intercept $G_{1A} + RT \ln h_{1A(2)}$). (After Powell, 1978.)

sites of substitution; that is commonly referred to in recent literature as the "Henry's law problem" (Fig. 2; Navrotsky, 1977).

In igneous petrogenesis

With reference to the geochemical modeling of partial melting and fractional crystallization in igneous petrogenesis, the simplest form of the

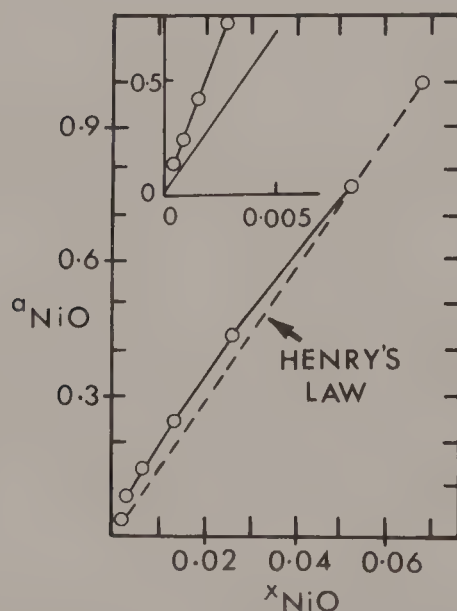


FIGURE 2. A plot of a_{NiO} (activity of NiO) against X_{NiO} (mole fraction NiO) showing the deviation of NiO activity from Henry's law in a mix of NiO in $\text{Na}_2\text{O}-4\text{B}_2\text{O}_3$ at $1,000^\circ\text{C}$. (After Navrotsky, 1978.)

equilibrium partition coefficient for an element (i) between two phases, mineral (α) and liquid (l) might take the form $D_{\alpha/l}^i$, which is the ratio of the concentration of the element in the two phases, i.e., concentration in mineral/concentration in liquid = D . In practice, partition coefficients can be measured directly from the run products of synthetic experiments (e.g., Takashashi, 1978), or from phenocrysts and matrix relationships in vitreous volcanic rocks. Measurements of chemical compositions may be acquired using an electron microprobe (e.g., Halden and Bowes, 1984), or the phenocrysts may be mechanically separated from the glassy matrix (e.g., Griffin and Rama Murthy, 1969) prior to wet chemical determinations on both the matrix and the phenocryst. Neither method of analysis takes into account the errors that can be introduced as a result of zoning within the phenocryst; only careful microprobe analysis, perhaps accompanied by an optical imaging by laser interferometry (to elucidate the complexity of zoning), may surmount this problem.

Some practical examples of the use of partition coefficients in igneous petrogenesis are summarized by Shaw (1979), who models Rayleigh fractionation (Wood and Fraser, 1976), accumulated Rayleigh fractionation, and batch melting for modal and nonmodal melting.

If single element-phase partition coefficients are used, then the effect of single-phase melting on restite and derived melt chemistry can be modeled. However, phase relationships for most rocks undergoing anatexis require that more than one phase enters the melt, necessitating the use of a bulk partition coefficient (D) of a given trace element for residual minerals at the time of separation of a liquid phase. ($D = D^i X^i$, where D^i is the partition coefficient for a trace element for a given phase, and X^i is the weight fraction of the phase in the solid).

For modal and nonmodal melting involving either Rayleigh fractionation or accumulated Rayleigh fractionation, theoretical considerations of these processes dictate that an infinitesimally small mass of liquid is separated from the residual solid and is prevented from any further reequilibration with the residual solid. The degree to which this can reflect real geological situations depends upon the distances that the trace element must diffuse to get into the melt, and the time that is taken to diffuse. Both distance and time are likely to be governed by crystal-lattice effects in the melting phase, and the degree of liquid reequilibration with the solid may be controlled by any polymerization within the melt. With batch melting, a mass of partially melted liquid is allowed to accumulate, but in this case is

assumed to have been in continuous equilibrium with the residual solid.

In the case of nonmodal melting, the proportions of the phases in the melt need not be the same as those in the residual solid; therefore knowledge of the bulk liquid partition coefficient (P) is required: $P = p^i D^i$, where p^i is the proportion of phase i entering the melt (D^i —see above). The inclusion of P modifies the modeling technique by the introduction of limiting factors. When a phase is exhausted in the solid, it cannot contribute to the chemistry of the liquid; here $F_{\max}$ (the maximum degree of partial melting) cannot exceed X^i/p^i —this is a mathematical limitation and does not preclude “geologically complete” anatexis. In addition, during nonmodal melting the bulk partition coefficient can no longer be assumed to be constant but is given by $D^n = (D - FP)/(1 - F)$ (where D^n is the changing bulk partition coefficient).

The effects of fractional crystallization on trace-element concentrations in an evolving melt can be modeled using $C_L/C_0 = F^{(D-1)}$. If the process is governed by Rayleigh fractionation, then an infinitesimally small amount of crystals precipitate from the liquid and are then prevented from any further re-equilibration with the liquid (perhaps achieved by cumulus phases settling out of a melt and subsequently being covered by other crystals). Whether this may be geologically reasonable must be argued on textural grounds, and is obviously untenable in cases where crystal resorption is evident. It is possible to represent this process graphically by drawing crystal fractionation vectors; if a bulk value of D is used, then the bulk evolution of the liquid is plotted. If an individual partition coefficient is used, then the potential effect of a single crystallizing phase (perhaps as phenocrysts) may be isolated. It is only where there is clear evidence of phenocrysts that this may be an accurate assumption; phase relations in most rocks may dictate that more than one phase would be crystallizing at any one time.

In practice, values for partition coefficients vary and are compositionally dependent (Takahashi, 1978; Cox et al., 1979, p. 334). It is also possible to relate the behavior of a trace element to those major elements that it will substitute for, leading to compound partition coefficients, e.g., for Ni/Mg and Sr/Ca. This approach can also be used to investigate the behavior of REE and trace elements in the presence of volatile elements such as chlorine, fluorine, and boron.

In metamorphic petrogenesis

The use of “partition coefficients” or, strictly speaking, equilibrium constants (K_D values) in

metamorphic petrogenesis for geothermometry and geobarometry is slightly different. Having the same basis in thermodynamic theory as partition coefficients in igneous petrogenesis, substitution in this case is governed by Raoult’s law, where there is only partial dilution of a solid-solution end member by another. Determinations of mineral chemistries are achieved in a similar manner to that applied to igneous rocks and are subject to the same error considerations. When a metamorphic reaction is investigated, considerable effort must be made to define the equilibrium assemblage, which in many instances is done solely on textural grounds.

The use of equilibrium mineral assemblages as geothermometers and geobarometers is a function of the dependence of a K_D upon temperatures and/or pressures of crystallization. In P – T space, if the partial derivative of $\ln K_D$ with respect to T at constant P is greater than the partial derivative of $\ln K_D$ with respect to P at constant T , then K_D depends more upon temperature than on pressure, and the equilibrium mineral assemblage under investigation will make a good geothermometer. If the opposite holds, then it will make a good geobarometer (Fig. 3; Wood and Fraser, 1976, p. 129).

The partial derivatives of $\ln K_D$ at constant P and T are rarely if ever zero, meaning that for an individual assemblage P and T cannot be determined independently. Geologically, this

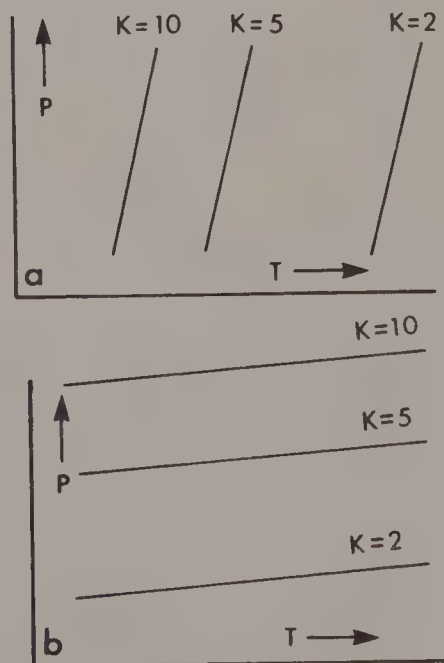
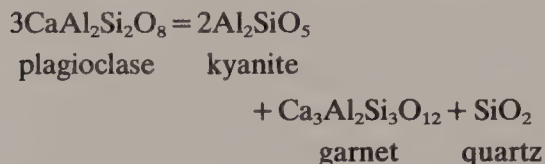


FIGURE 3. Curves representing constant K_D in P – T space, (a) for a potential geothermometer and (b) for a potential geobarometer. (After Wood and Fraser, 1976.)

problem may be resolved by using different equilibrium mineral assemblages from different lithologies that various lines of evidence indicate to have been subject to the same conditions of metamorphism.

Considering as a practical example a potential geobarometer using the mineral reaction



and assuming that the assemblage is in equilibrium, then

$$\Delta\mu = 0 = \Delta G^\circ + RT \ln K_D.$$

From the basic thermodynamic data of the phases for entropy, enthalpy, and volume changes involved in the reaction, it is possible to calculate the formula of a line in P - T space:

$$RT \ln K_D = 6.62P - 0.1548T + 50.6$$

where temperature is in kelvin and pressure in kb (Powell, 1978, p. 244). This geobarometer is developed further by Newton and Haselton (1981); K_D is calculated as a ratio of chemical activity of grossular in the garnet to anorthite in the plagioclase. The chemical activities are determined as complex functions of the mole fractions of anorthite and grossular in the respective phases; electron microprobe analysis can provide the basic data for such calculations.

If the assemblage contains coexisting garnet-biotite-cordierite, then it may be possible to determine temperatures of crystallization based on the distribution of Fe and Mg (and Mn) between these phases; examples of these geothermometers are given by Perchuk et al. (1981). Given that the assemblage can provide both pressures and temperatures of crystallization, complex metamorphic histories can be elucidated (Halden and Bowes, 1984).

Controls

The distribution of major and trace elements between phases during igneous and metamorphic processes can be controlled by pressure, temperature, ionic radius, coordination within crystal lattices, and charge (e.g., Eu distribution can be influenced by f_{O_2}). Therefore, to make most efficient use of partition coefficients and equilibrium distribution constants in the solution of petrogenetic problems, complete knowledge of the petrography, major- and trace-element chemistry, volatile fugacities and geological setting of the rock are required.

N. M. HALDEN

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Cross-references: *Anatexis; Metamorphic petrology—use of thermodynamics; Metamorphism.*

ELEMENTS—See Vol. IVA

ELEMENTS : PLANETARY ABUNDANCES AND DISTRIBUTION—See Vol. IVA

EMPLACEMENT—MODES

The term emplacement used in the context of petrology and structure means the attainment of the present position of a particular intrusive mass of rock. It is most commonly used in association with intrusive igneous rocks, but it is also used in relation to the modes of emplacement of material (other than magma) infilling volcanic necks and breccia pipes, as well as to salt domes and other intrusive sediments. Emplacement may also be used in connection with the conversion in situ of crustal rocks into rocks of igneous aspect (feldspathization-granitization), but this use of the term is excluded from this description.

Magmas

Magmas move in the crust owing to the influence of gravity, tectonic forces, or pressures from within the magmas themselves. Their movement necessitates displacements of the surrounding crustal rocks and accordingly their mode of emplacement is often controlled by structures within these rocks.

Forceful intrusions. *Forceful (forcible)* intrusions are those in which the magmas make room for themselves by pushing aside the crustal rocks. Magmas will tend to follow planes of least resistance in the rocks they are penetrating. If these planes are flat-lying (e.g., bedding planes) various types of concordant intrusions will be produced. *Sills* are formed where magmatic pressures are sufficient to part the rocks along bedding planes and allow emplacement of sheets of magma (Fig. 1a) (see **Igneous intrusions—structural behavior**). Where pressures are sufficient to part the rocks along bedding planes and to overcome the weight of overlying crustal rocks with the arching up of the roof, *laccoliths* are formed: these intrusions have a planar floor and gently convex roof (Fig. 1c). Where the pressure of a rising column of magma arches the overlying crustal rocks by a

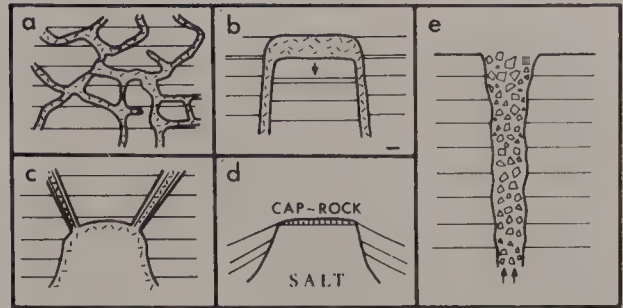


FIGURE 2. (a) Vein network, (b) ring dyke, (c) cone sheet, (d) salt dome, (e) breccia pipe.

process of stretching and minor fracturing, a *dome* is formed (Fig. 1d).

In some cases, the planes of least resistance may run across the bedding-planes in the crustal rocks (e.g., joints, faults) and magmas forced along them will produce various types of discordant intrusions. A *dyke* is formed where magma is forced upwards along fracture planes to produce a vertical or steeply-inclined sheet of intrusive igneous rock (Fig. 1b). Where, after the development of a dome, and with increasing pressure, the magma forcibly breaks through the overlying crustal rocks, *diapiric intrusions* are formed (Fig. 1e); for example, many stocks and bosses of granite have formed in this way (see **Diapirism**), and on a larger scale, forcible intrusion of granitic magma appears to have been the most important mode of emplacement of the Sierra Nevada Batholith (Bateman et al., 1963).

Passive intrusions. *Passive (permitted)* intrusions are those in which the magma rises into the crustal rocks to occupy spaces usually caused by subsidence or faulting. These intrusions are generally discordant. In the formation of vein networks, the magma follows a ramifying fracture system in crustal rocks previously broken up by explosive activity or faulting (Fig. 2a). Where, at very shallow depths, magma rises into open fissures, dykes result (Fig. 1b). Where the magma rises along approximately vertical concentric fractures developed over a subsiding block of crustal rocks, *ring dykes* are produced (Fig. 2b). Where it rises along conical concentric fractures inclined towards an underlying dome-shaped mass of magma, *cone sheets* are produced (Fig. 2c).

Magmatic stopping. Some magmas have been emplaced by detaching blocks of the country rock, which foundered into the rising magma (Fig. 1f). This process of magmatic stopping was first proposed by Daly (1933) and has been used by others, including M. P. Billings (see reference in Badgley, 1965, chap. 9) to explain the emplacement of some stocks.

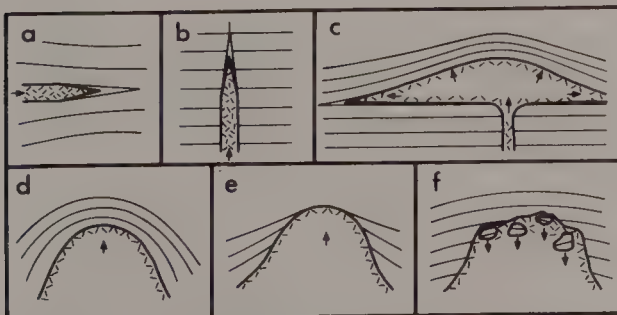


FIGURE 1. (a) Sill, (b) dyke, (c) laccolith, (d) dome, (e) diapiric intrusion, (f) pluton emplaced by magmatic stopping.

One of the most important aspects of magmatic stoping is the subsidence of blocks of crustal rocks. This must be distinguished from the carrying upwards of detached blocks by a forceful rising magma which is often accompanied by flow structure in the intrusive rock. The formation of a ring dyke involves the subsidence of a large block of crustal rocks and is essentially a type of magmatic stoping (Fig. 2b). This process can also be involved in the other modes of emplacement, for example in the latter stages of the emplacement of any dome-like forceful intrusion, subsidence may take place that involves magmatic stoping on any scale and perhaps also cauldron subsidence at the Earth's surface.

Other factors influencing the modes of emplacement of magmas. The time of emplacement relative to orogeny (pre-, syn-, or post-tectonic), the depth at which emplacement takes place, the physical condition of the crustal rocks, and the physical condition of the magma all influence the mobility of a magma within the crust. A magma may still remain mobile even when partly crystallized, and if intruded in this stage, a flow structure will be preserved in the resulting igneous rock. The mobility of a body of magma normally ceases after consolidation, but later tectonic movements may partly remobilize the intrusion and make it capable of further movement.

These factors and many of the modes of emplacement are dealt with in Newall and Rast (1970), where an emphasis is placed on the physical conditions of initiation, ascent and emplacement of magmas at various levels of the Earth's mantle and crust. In his description of an Andean batholith, Pitcher (1978) includes many forms of intrusions and demonstrates that ring complexes form direct connections between intrusions of the batholith and the volcanic masses of the country rock and cover.

Fragmentary rocks in pipes

Instead of containing solidified magma, some pipes and fissures feeding volcanoes may be filled with fragmentary rocks (Fig. 2e—see **Breccia pipe**). The individual fragments contained within these rocks may be of lava or of crustal rocks through which the pipe or fissure passes. Other pipes, which may never have reached a contemporary land surface, are infilled with similar breccias but contain, in addition, fragments of deep crustal or subcrustal rocks (see **Kimberlite**). The pipes and fissures were probably cored out by gases emanating from an underlying magma, and the breccias illustrate emplacement of material by upward-streaming gases (Fig. 2e) (see **Fluidization**).

Intrusive sediments

Salt. With the growing thickness of accumulating sediments on top of salt beds, the pressure increases and the salt reacts differently from most other rocks in that it begins to flow rather than fold or fracture. This may result in the salt bulging upwards, doming the overlying sediments and eventually forcibly breaking through these sediments to form diapiric salt domes (Fig. 2d).

Mudstones. Some mudstones may react similarly to salt and produce diapiric dome structures (see examples in Holmes, 1978, chap. 10).

Sandstone. For some time after deposition sands remain water-saturated and capable of mobilization. If these sands are subjected to pressure of overlying sediments, they may be forcibly injected as concordant sheets (sills) into the surrounding sediments or as discordant sheets (dykes) into dilational cracks or fissures in these sediments. Sandstone dyke injection may be upwards from below or downward from above (see examples in Collinson and Thompson, 1982, chap. 9).

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Cross-references: *Batholith; Breccia pipe; Calderas and volcano-tectonic depressions; Circum-Pacific plutonic and volcanic activity; Cone sheets; Diapirism; Diatreme; Dyke (dike); Fluidization; Igneous intrusions—structural behavior; Intrusion; Kimberlite; Ring dyke; Sill.*

ENCLAVE

An *enclave* is a piece of rock totally enclosed within, and differing in some way from the fairly homogeneous igneous rock in which it is found.

The term (Lacroix, 1890) does not carry the genetic connotation of the terms xenolith (Sollas, 1894) and autolith (Holland, 1900), which imply foreign and cogenetic origins, respectively. The term inclusion is also used, but this can be applied as well to particles included in single crystals.

Enclaves offer important information relating to the formation and emplacement of the granitic masses that contain them (Didier, 1973). Evidence from their study points to (1) an anatectic origin of autochthonous migmatitic granites, and of the magma of intrusive leucogranites; (2) a mixed origin for common granodiorites and monzogranites in orogenic belts (a possible genetic model is the melting of the deep crust around basic intrusions and the imperfect mixing of the two magmas); and (3) a limited role for assimilation or granitization of host rocks.

Xenoliths are fragments of sedimentary, igneous, or metamorphic rocks that have been incorporated in a melt, generally granitic, and that have suffered contact metamorphism (Fig. 1a,c). Their earlier textures and structures have generally not been obliterated completely by recrystallization and they are common in a wide range of intrusive masses. In some xenoliths of hornfels the presence of granitic "bands" and veins indicates the beginning of melting (contact anatexis). The presence of feldspathic megacrysts in the xenoliths in porphyritic granites has been used by some workers as indicative of a metasomatic origin of the granite (see *Xenolith*).

Surmicaceous enclaves (Lacroix, 1933) are not very common in intrusive granitic masses but are abundant in autochthonous granites related to migmatites, where they coexist with schist fragments that represent different stages of melting and transformation into surmicaceous enclaves. They have a lenticular form and are characterized by abundance of mica; biotite occurs alone or with muscovite, according to the nature of the host granitic rock. Al-rich minerals such as andalusite, sillimanite, cordierite, corundum, and spinel are also very abundant; sulfides and oxides are common.

These enclaves are restites (refractory residues remaining after fusion). In some cases they appear to be of deep-seated origin cogenetic with their enclosing granite. In other cases they result from the fusion of pelitic hornfels fragments by contact anatexis.

Microgranular enclaves (Didier and Roques, 1959; Didier, 1973) are composed of microgranular igneous rocks, though the term is used to include fine-grained rocks (Fig. 1b,c). Composition varies within wide limits; some are essentially the same as the host rock with only

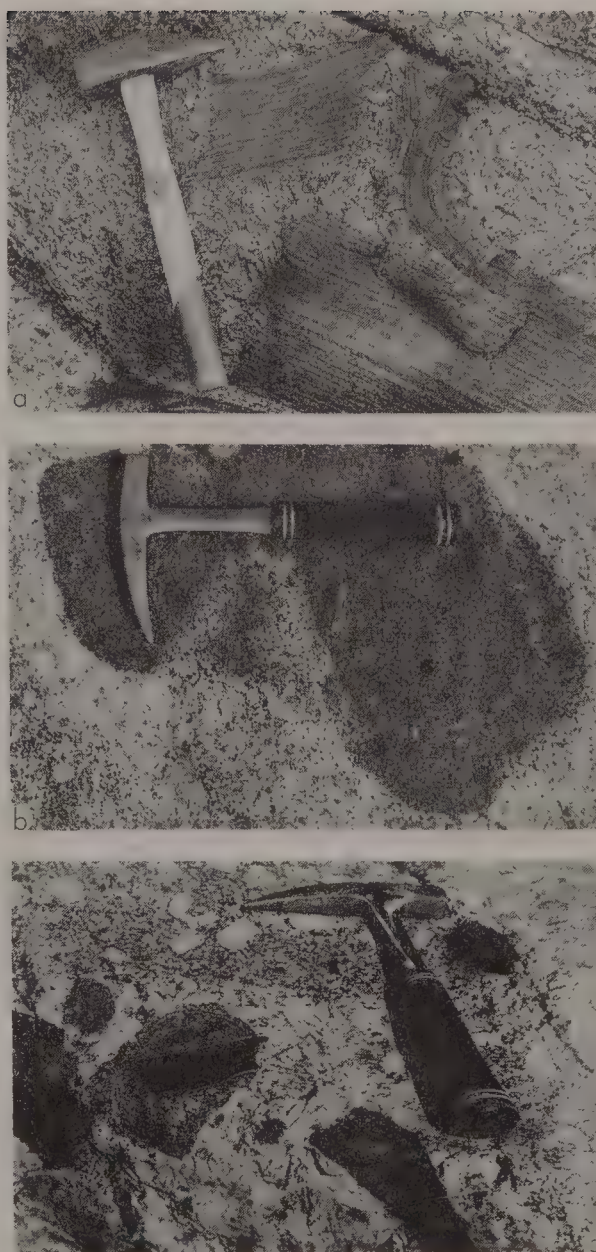


FIGURE 1. (a) Hornfels enclave (xenolith) in the Ploumanac'h granite, Brittany, France. (b) Microgranular enclave with granitic veins; Sidobre granodiorite, Massif Central, France. (c) A dark microgranular enclave within a hornfels enclave (xenolith); the contact is emphasized by a thin quartzofeldspathic fringe; St. Guival-Liron granodiorite, Massif Central, France.

textural differences, while others have strikingly different compositions, generally considerably more basic. Their texture may be porphyritic, aplitic, or doleritic. There is no evidence of recrystallization and this feature provides a means of distinguishing them from xenolithic fragments associated with the emplacement of the host granite that show variable stages of change to hornfels.

The composition of the lighter-colored enclaves is similar to that of the host granite and they may represent early-crystallized fractions

of the granitic magma. Many may represent a disrupted early border facies and they are particularly abundant in the high-level intrusions, where they may attain large dimensions.

The darker microgranular enclaves are essentially localized in granodiorites and monzonites but some are found in the alkaline granites and syenites of subvolcanic complexes. They are absent in leucogranites and autochthonous granites. Both textural and compositional features indicate derivation from basic rather than granitic rocks and they most commonly occur in large masses of basic rocks. They have been interpreted variously as restite, cumulates or products of imperfect mixing of felsic and basic magmas. The latter interpretation is supported by a comparison with enclaves of intermediate lavas, which have many similar petrographic characters (chilled margins, crenulated margins, incorporation of xenocrysts giving ocellar quartz or rapakivi feldspars). It is also supported by the nature of the contact between granitoid masses and some large basic masses where microgranular enclaves occur as products of hybridization, chilling, and disruption of the basic magma.

The regular association of basic enclaves with granodiorites and monzonites suggests that injection of basic material of mantle origin into the deeper parts of the crust could be the factor responsible for the formation of granodioritic magma by anatexis.

Metamorphic enclaves from autochthonous granitic massifs linked to migmatites are very common. Most are fragments of schist in different stages of transformation into surmicaceous enclaves with partial fusion having resulted in an impoverishment in quartz and feldspars but an enrichment in biotite. By contrast with those in the intrusive granitic massifs, enclaves in migmatitic massifs contain few Al-rich minerals like andalusite, sillimanite, corundum, and spinel. Most fragments are refractory rocks such as amphibolite, pyroxenite, quartzite, and recrystallized older igneous rocks.

J. DIDIER

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Cross-references: *Anatexis; Assimilation; Batholith; Emplacement—modes; Granite; Hornfels; Magma; Migmatite; Xenolith.*

ENDOGENOUS DOME

The term *endogenous dome* (cupolar di ristagno, dome endogene, staukuppe) indicates a volcanic edifice made up by lava which, accumulated above the volcanic vent, formed a more or less rounded mass owing to slow successive extrusions. Each inflow intrudes beneath the previous one, so that the innermost part of the dome is formed by the latest supply of magma (Fig. 1). In most cases, the successive inflow of magma beneath the already cooled carapace of the dome creates tensional fractures that may be healed by further magma emplacement along them. In some cases there are effusive phases. The term endogenous dome is also used for the more complex edifices in order to distinguish them from exogeneous domes due to the outflow of less viscous magmas.

Endogenous domes are formed only by highly viscous melts of salic composition (rhyolites, dacites, andesites, trachytes, and phonolites). However, as the viscosity is not only a function of the chemical composition of the magma, but also of the temperature and of the gas content, more femic magmas may be sufficiently viscous to form domes if they are relatively cool and degassed and laden with phenocrysts.

The domes consist of two parts that show

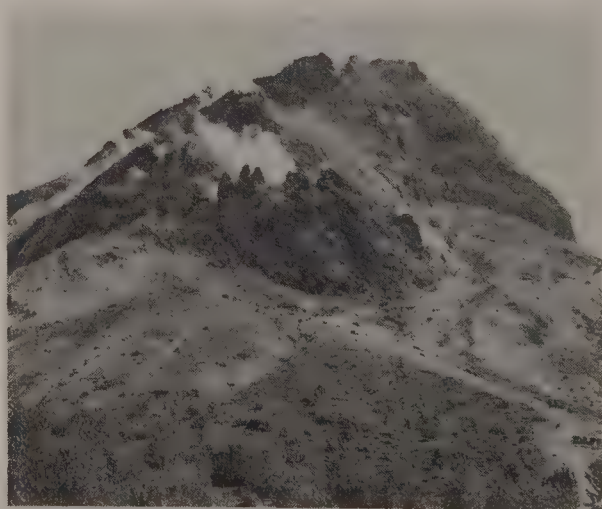


FIGURE 1. Dacitic endogenous dome on the eastern slope of the volcano Usu, Syowa-Sinzan, Japan.

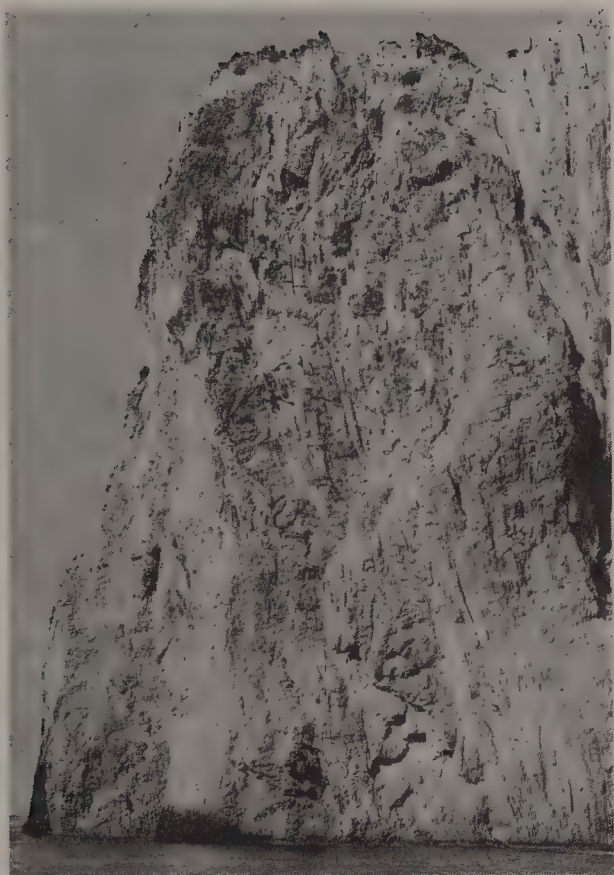


FIGURE 2. Laminar flow in rhyolitic endogenous dome, Palmarola, Pontine Islands, Italy.

different structures: a more or less compact inner part, and an outer part that is formed by blocks and detritus produced during the growth of the dome and which rolls down the slope. This material accumulates around the dome, forming a scree (talus slope) that may reach considerable thickness and cover a great deal of the edifice.

In nearly all endogenous domes, the inner part shows a fluidal structure with flow lines that are vertical in the innermost parts (Fig. 2) and progressively become flatly disposed outwards.

The form of the domes is often very irregular, with the presence of protrusions in the form of pinnacles, towers, and needles. In a few cases, such protrusions reach considerable dimensions and even become the predominant feature of the dome. The appearance of such nearly solid protrusions generally marks the final phase of the eruptive cycle in which the magma has become extremely viscous by cooling and degassing. Most of the big protrusions are only short-lived because, in consequence of the shrinkage by cooling, they crumble to blocks and collapse (e.g., endogenous domes with spines at Tarumai, Japan; Mt. Pelée, Martinique; and Santa Maria, Guatemala).

Direct observation of dome growth has been

made at Georgios, Greece (1866–70); Mt. Pelée, Martinique (1902–4); Tarumai, Japan (1909); Santa Maria, Guatemala (1922–34); Dafni, Greece (1925–26); Merapi, Java (1883–1930); and Syowa-Sinzan, Japan (1943–45). These observations indicate that explosive phases play an important role not only before the initiation of the dome but also associated with opening the vent along which the magma rises. Intermittent strong explosions also occur, resulting in the emission of glowing clouds on the flanks of the growing domes. These suspensions of glowing lava fragments and shards in very hot gases move very rapidly down the slope of the volcanic edifice, sometimes covering wide areas in which all life is destroyed (e.g., Mt. St. Helens, Oregon, U.S.A., 1980—see Decker and Decker, 1981).

Sometimes violent explosions also mark the end of the eruptive cycle. Such explosions may cause the partial or even total destruction of the dome itself, as in the cases of Dafni (1926) and Merapi (1934).

L. VILLARI

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Cross-references: *Acid(ic) rocks*; *Banding in igneous rocks*; *Lavas and lava eruptions*; *Volcanoes and volcanism*.

EPIDOTE AMPHIBOLITE FACIES

The *epidote amphibolite facies* is defined in basic rocks by the assemblage albite + epidote + hornblende. It is distinguished from the greenschist facies by the presence of hornblende rather than actinolite, and from the amphibolite facies by the presence of sodic plagioclase + epidote rather than more calcic plagioclase. Almandine garnet is typical of pelitic rocks of this facies and can also occur in basic rocks. This facies is regarded by many as a subfacies of the greenschist facies or a transition from greenschist to amphibolite (Winkler, 1967; Turner, 1981). The boundaries are not accurately defined by experimental studies. At lower grades, the boundary is marked by the change from actinolite to hornblende and with increasing temperature by the disappearance of epidote

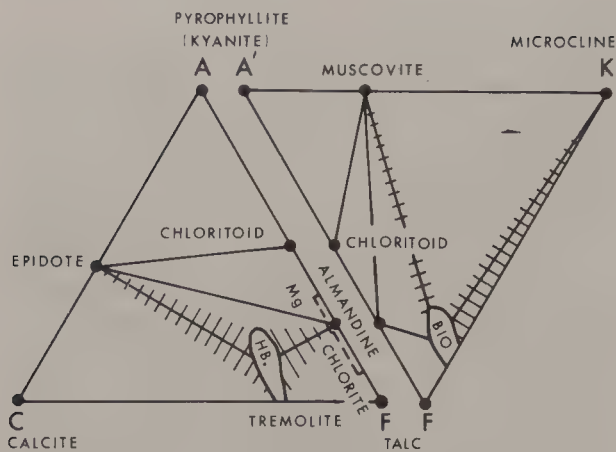


FIGURE 1. ACF and A'KF diagrams of the epidote amphibolite facies; cross-hatching indicates variability in chemical composition of hornblende, chlorite and biotite. (After Winkler, 1967.)

and a sharp jump in the anorthite composition of plagioclase. At higher pressures, the epidote amphibolite facies lies adjacent to the eclogite facies. There are examples known of eclogites occurring in close association with epidote amphibolite facies rocks. However, they are normally regarded as forming in dry rocks but under the same *P-T* conditions as the hydrous epidote amphibolite assemblages around them (Morgan, 1970). The progressive transition has not been recorded in nature.

Basic rocks. As the facies is defined on the basis of basic rocks, there is little variation in the mineralogical composition of basic rocks. Some may contain almandine (Fig. 1).

Pelitic rocks. The epidote amphibolite facies in basic rocks roughly corresponds to the almandine garnet zone of the classic Barrovian sequence in pelitic rocks. Chlorite is only stable

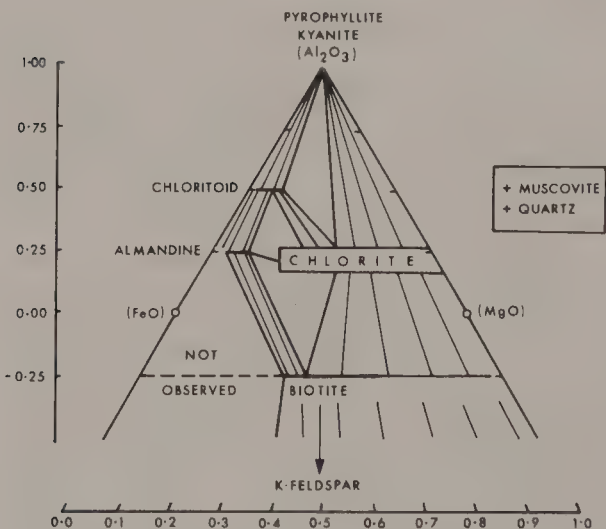


FIGURE 2. AFM projection of pelitic rocks of the epidote amphibolite facies. (After Winkler, 1967.)

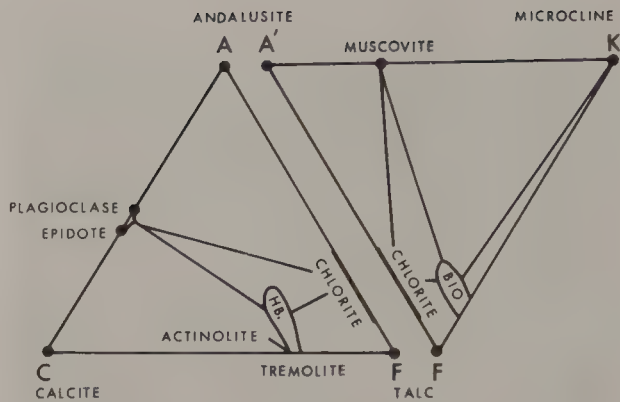


FIGURE 3. ACF and A'KF diagrams of the quartz-andalusite-plagioclase-chlorite subfacies. (After Winkler, 1967.)

in magnesian rocks; mostly pelitic schists are garnet-biotite-muscovite rocks (Fig. 2). Chloritoid is most commonly found in Fe-rich pelites. The upper boundary of pyrophyllite changing to kyanite probably also occurs in this zone and high-grade pelitic rocks of the Alpine-type facies series (jadeite-glaucophane type) may contain both glaucophane and kyanite. Winkler (1967) suggests that andalusite is stable at low pressures in this facies rather than pyrophyllite (Fig. 3).

Calcareous rocks. Probably no diagnostic differences occur between impure calcareous rocks in the greenschist and epidote amphibolite facies—unless they are very impure, when hornblende becomes stable in preference to actinolite. Epidote and tremolite are stable in both facies in calcareous rocks. Dolomite, although not common in this facies, is probably unstable in higher grades of the greenschist facies also.

Ultramafic rocks. There are only slight differences between the parageneses found in the greenschist facies and those of the epidote amphibolite facies. Forsterite is stable throughout this facies. Miyashiro (1973) suggests that anthophyllite is a stable ferromagnesian mineral in this facies, but according to Evans and Trommsdorff (1970) it is not stable in ultramafic rocks until temperatures of about 670°C.

Other minerals stable at this facies in ultramafic rocks are antigorite, talc, diopside, tremolite, chlorite, and magnesite. Magnesite, dolomite, and calcite are frequently formed on introduction of CO₂.

A. E. WRIGHT

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Cross-references: *Amphibolite facies*; *Deserpentinization*; *Glaucophane schist facies*; *Greenschist facies*; *Marble*; *Metamorphic facies series*; *Regional metamorphism*.

EUTECTIC MELTING

Solid-liquid equilibria in chemical systems exhibit two general types of relationships, the eutectic type and the solid-solution type; there are all gradations between the two. In silicate systems, liquids are usually completely miscible. However, in eutectic systems, the solid phases are immiscible or only partially miscible. The *eutectic* is the lowest melting point of the system under its own pressure; the corresponding temperature is the *eutectic temperature*, and the liquid formed at the eutectic is the *eutectic liquid*.

Eutectic liquids lie between the solid phases of the system in composition. They are congruently saturated with all the solids, since the addition of heat causes direct melting of some of each solid phase to form the liquid; removal of heat causes precipitation of all solid phases. Depending upon the number of components in the system, binary eutectics, ternary eutectics, quaternary eutectics, etc., can exist. A rock whose minerals are almost completely in eutectic proportions can be referred to as being *anchieutectic*.

Phase diagrams

Figure 1 shows a simple binary eutectic system (A–B) with both temperature and pressure as variables; there is no miscibility between the solid phases. Isotherms are shown on the two curved liquidus surfaces. These surfaces show the liquid compositions that coexist with crystals of pure A or B, at equilibrium, at temperatures between the melting points of the pure solids and the eutectic curve E–E₁. In the absence of volatiles, increasing pressure usually raises the melting points of the pure solids and the minimum melting temperature of any mixture of the solids (eutectic). In general, the composition of the eutectic liquid varies with pressure.

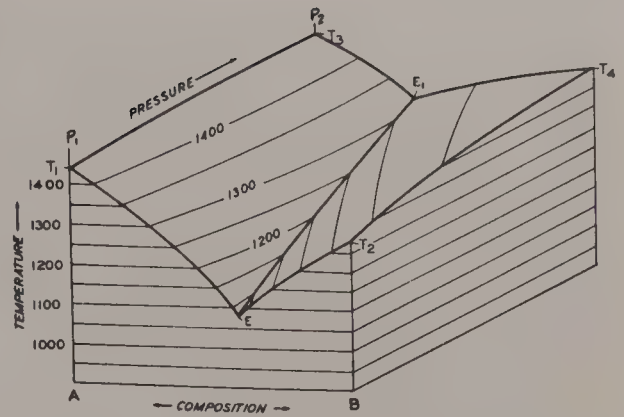


FIGURE 1. Simple binary eutectic system (A–B); temperature, pressure, and composition variables all shown. Isotherms are shown on curved liquidus surfaces; T_1 , T_2 , etc., are melting points of pure A or B at pressures of P_1 and P_2 ; E–E₁ is the eutectic for the system. (From Wahlstrom, 1950.)

Figure 2 is a constant-pressure (1 atm) section of the two-component silicate system diopside–anorthite. The two liquidus curves start at the melting points of pure diopside and pure anorthite, and intersect at the eutectic E. Liquids on a liquidus coexist with either diopside or anorthite over the temperature range in which both liquid and crystals can be present. Only at the eutectic can liquid coexist with both crystal phases. A liquidus is a solubility, or saturation, curve showing the amount of a component dissolved in the liquid at a given temperature when crystals are also present. At constant pressure, the phase rule

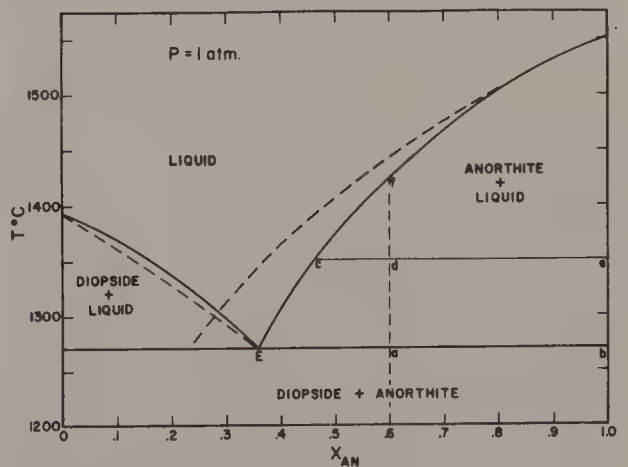


FIGURE 2. The eutectic silicate system diopside–anorthite. Solid-liquidus curves are experimentally determined; dashed curves are calculated assuming ideal liquid solutions and constant values for ΔH and ΔV . Horizontal line at 1270°C is the solidus; c–e is a tie line connecting the compositions of coexisting phases at 1350°C.

shows that the system is univariant on the liquidus; the system is invariant at the eutectic.

Below the eutectic no melting occurs. When the temperature increases to the value of the eutectic temperature, melting begins in any mixture of diopside and anorthite, regardless of their mixing proportions. Experimental evidence shows that such melting is to be expected at the contacts between adjacent crystals of diopside and anorthite. This melting temperature, below those of the pure solid phases, can be explained in terms of the disrupting influence of one crystal structure on the other. The composition of the first liquid formed is given by E (eutectic liquid). Further addition of heat to the system causes additional melting of the two solids. Both solids melt simultaneously in proportions given by E, such that the liquid composition remains unchanged. During eutectic melting the temperature of the system remains constant at E; the added heat is balanced by the latent heats of fusion and is contained within the liquid phase owing to its higher heat capacity.

A solid mixture with the same proportions of crystal as the eutectic composition will melt completely at the eutectic, yielding a liquid of that composition. In all other mixtures, the process of melting at the eutectic will continue until one of the solid phases is consumed. In such mixtures, the system will then be composed of either diopside crystals plus liquid or anorthite crystals plus liquid, depending upon whether the initial crystal mixture was to the left or right of the eutectic composition, respectively.

With further addition of heat, eutectic melting ceases and the system follows one of the liquidus curves. The liquid composition changes and the temperature of the system increases as the remaining crystal phase continues to melt. Upon complete melting, the liquid will have achieved a composition identical to that of the initial total crystal mixture. Further addition of heat will then raise the temperature of this liquid.

The relative proportions of solid and liquid phases can be determined throughout the melting process by geometrical constructions on the phase diagram. Actual amounts of phases can be determined if the amounts of the initial crystals are known. In the example construction in Figure 2, the initial mixture contains 40% diopside and 60% anorthite. When eutectic melting has consumed all diopside, segments of the tie line E-b give the relative proportions of liquid and remaining anorthite; liquid/anorthite = ab/Ea and the eutectic liquid composition is 36% anorthite plus 64% diopside. Proportions of the three phases present during eutectic

melting, before diopside is consumed, can be determined numerically by stating the proportion of liquid present and subtracting this amount of eutectic liquid composition from the initial bulk composition. Proportions of phases present above the eutectic temperature are given by tie lines such as c-e. At 1350°C, liquid/anorthite = de/cd and the liquid contains 46% anorthite and 54% diopside. At the temperature of f, the system is completely melted.

Thermodynamic considerations

The thermodynamics of eutectic melting is most easily stated for systems with complete immiscibility between the solid phases; pure solids then coexist with the eutectic liquid. The Van't Hoff equation can be used to derive an expression for the coexistence (at equilibrium) of a pure solid component (A) and that component in a liquid solution. Adding the effect of pressure gives

$$\ln X_A^L \gamma_A^L = \int_{T_A}^T \frac{\Delta H_A}{RT^2} dT - \frac{1}{RT} \times \int_{P^0}^P \Delta V_A dP$$

where

X_A^L = mole fraction of A in liquid,

γ_A^L = activity coefficient of A in liquid,

ΔH_A = latent heat of fusion of A at T,

ΔV_A = change in volume of A upon fusion, and

T_A^0 = melting point of pure solid A at standard pressure P^0 .

The first term on the right gives the effect of temperature on the solubility of A in the solution, while the second term gives the effect of pressure. In general, ΔH and ΔV will vary with temperature and pressure and appropriate expressions showing these dependences must be substituted into the equation. For many purposes, and especially for small changes in T and P, ΔH and ΔV can be considered as constant. Further, if the solution is perfect, the activity coefficient becomes unity. Since X_A^L is the solubility of A in the solution by definition, the above expression defines the solubility surface over a T and P range or the solubility curve at constant P. In a binary system, two such curves are present in the phase diagram (Fig. 2); the second curve is given by

$$\ln(1 - X_A^L) \gamma_2^L = \int_{T_B}^T \frac{\Delta H_B}{RT^2} dT - \frac{1}{RT} \int_{P^0}^P \Delta V_B dP.$$

The eutectic is at the intersection of the two curves and can be determined by the simultaneous solution of the two equations for freezing-point depression.

BERT E. NORDLIE

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Cross-references: *Magma; Phase rule; Silicate melting.*

melting reactions, including solid-liquid reactions and solid-liquid-vapor reactions, have been studied and solubility measurements have been carried out, including investigations of the solubility of water in silicate melts and the solubility of solids in an aqueous gas phase. Rock-melting studies (including quenching in order to preserve certain physical-chemical characteristic of the high-temperature state that would be changed by slow cooling) have played a major role in the study of magma generation and also of the nature of the Earth's crust and mantle.

The range of experimental petrology and of techniques employed are given by Ulmer (1971) and Edgar (1973). The application to petrogenesis of this branch of petrology is illustrated by the work of Bowen (1928), Fyfe et al. (1958), Tuttle and Bowen (1958), Yoder and Tilley (1962), Wyllie (1971), Ringwood (1975), Vernon (1976), Yoder (1979), Morse (1980), Turner (1981), Maaløe (1985), and Edgar (1987), and in many articles (and references cited) in this volume.

D. R. BOWES

EXPERIMENTAL PETROLOGY

Experimental petrology is concerned with experiments designed to reproduce in the laboratory conditions of formation of minerals and rocks, with the aim of elucidating the nature of petrological processes. The experiments deal with the physical properties or physical chemistry of minerals, rocks, and rock melts and of solutions, gases, and vapors coexisting with solid or molten materials. The pioneer work was done in Edinburgh, Scotland, at the beginning of the 19th century by Sir James Hall on rocks and minerals sealed in gun barrels at high pressures and temperatures. This initiated investigations of simple silicate systems, of metamorphic (re)crystallization, and of melting relations of rocks, aspects that gained tremendous impetus in 1904 when the Geophysical Laboratory was founded in Washington, D.C., U.S.A., and as new techniques were developed for the study of silicate melts at carefully controlled temperatures (see the *Carnegie Institution of Washington Yearbook*). The design of equipment capable of maintaining high temperatures with high pressures mainly took place after 1945 and there are now many laboratories throughout the world engaged in experimental petrology.

The range of work covers the determination of physical properties of rocks and minerals such as elastic constants, strength, flow, fracture, electrical conductivity, and seismic velocity. Subsolidus reactions, including solid-solid transitions and dissociation reactions, and

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Cross-references (selected): *Anatexis*; *Basalt*; *Element partitioning in petrogenesis*; *Eutectic melting*; *Granite*; *Lunar petrology*; *Metamorphic petrology—use of thermodynamics*; *Metamorphic reactions*; *Migmatite—origin of neosome*; *Nucleation of minerals in magmas*; *Phase transformation*; *Silicate melting*; *Upper mantle—experimental petrology*; *Water in rock melts*.

EXPLOSIVE VOLCANIC ERUPTIONS—CLASSIFICATION

Most volcanic eruptions are to some extent explosive, and the explosivity index, which is the percentage of pyroclastic material (i.e., volcanic material fragmented and ejected from the vent by volcanic explosions) in the total erupted products is often 100%. The *volcanic explosivity index* is an informal 0–8 scale of increasing explosivity based on erupted volume, plume height, duration of climactic phase, etc., intended as a useful measure of the likely impact of the eruption on man.

There are several different measures of eruption size: (1) magnitude—the total quantity of pyroclastic products (taken as the equivalent volume of dense rock or magma without voids), for which Tsuya's scale (magnitude 9— $>100 \text{ km}^3$, 8—10 to 100 km^3 , 7—1 to 10 km^3 , etc.) is a convenient measure, (2) intensity—the discharge rate of magma plus rock from the vent, and (3) violence—the extent to which dispersal is determined by the translational kinetic energy of the pyroclastic system.

The jet emitted from the vent in an explosive eruption is a mixture of gases and pyroclasts which slows as it rises. If the density of the mixture at the top of the jet is lower than the ambient atmosphere at that level, the mixture continues to rise as a convective plume, with the plume height depending on energy-output rate; if the density is greater, eruptive column collapse occurs.

Pyroclasts released from a plume settle out to form a pyroclastic fall deposit. Pyroclasts from a high plume form widely-spread layers; from a low column they build a cone localized immediately around the vent instead (their eruptions are loosely described as “sheet-forming” or “cone-building,” respectively). Column collapse generates pyroclastic flows and surges that flow across the landscape.

Terms such as strombolian, vulcanian, plinian, and peléan have long been informally applied to different styles of explosive activity.

A well-defined magmatic series ranges in order of increasing pyroclastic dispersal from *hawaiian* through *strombolian* and *subplinian* to *plinian*. Their deposits consist mainly of strongly vesicular juvenile material (scoria, pumice, shards) and tend to be poorly bedded because they come from a continuous gas blast operating out of an open vent. Spatter is the most distinctive pyroclastic product of hawaiian activity, cinder cones of strombolian activity, low-angle pumice cones of subplinian activity, and extensive pumice sheets of plinian activity.

Eruptions in which a significant amount of non-volcanic water participates are called *phreatomagmatic*. Interaction with water causes the magma to be strongly fragmented, and cohesion of damp ash particles causes premature ash deposition. Phreatomagmatic eruptions include the *surtseyan* type, which produces ash-rings localized around the vent, and the sheet-forming and generally more silicic *phreatoplinian* type.

Eruptions of *vulcanian* type are very common as discrete explosions on andesitic volcanoes, and generate dark clouds heavily laden with fine ash much of which may consist of dense rock debris. Opinions differ regarding the role of groundwater and the effect of the magma rheology (vulcanian eruptions commonly involve crystal-rich magmas). Much mechanical attrition occurs when ejecta thrown up by each explosion from a deep crater fall back into the crater. Often only the finer ash clears the crater rim.

Phreatic explosions are those in which magma supplies the energy but no magma is erupted; they often occur in lake-filled craters, and their ejecta often include much muddy (hydrothermally-altered) material. *Hydrothermal explosions* are discharges from geothermal fields and are not directly related to magmatic activity. Many phreatomagmatic, phreatic, or hydrothermal explosions give rise to *maar volcanoes*, characterized by a crater (often lake-filled) descending well below the general surface level and having a rather low rim.

Eruptions which generate *nuées ardentes*, the dangerous glowing avalanches which sweep down the sides of steep volcanic cones, are often called *peléan* after the catastrophic outburst of Mt. Pelée (Martinique) in 1902. Mechanisms include down-slope avalanching of fall-back ejecta from eruption-column collapse (St. Vincent type), outwardly-directed blast from an inclined vent (peléan type), or blast outwardly deflected by the topography, and the collapse and disintegration of viscous lava on a

steep slope (Merapi type). Glowing avalanches are pyroclastic flows and/or surges, and have an impressive over-riding ash cloud. The 18th May, 1980 directed blast from Mt. St. Helens was not incandescent.

Ignimbrite eruptions generate highly pumiceous pyroclastic flows; it is thought generally that they result from column collapse. Ash flows are highly mobile because the material in them is in a semi-fluidized condition, and may travel 10 to 100 km from the vent even where a steep cone is lacking. The resulting rock-body is called *ignimbrite* (or ash-flow tuff). Because of efficient heat conservation in the ash flow, the pyroclastic particles in many ignimbrites have united together after deposition to form welded tuff (not all welded tuffs, however, originate from ash-flows: some are of fall origin). Ignimbrites include the biggest single eruptive units known, having individual volumes exceeding 1,000 km³. Fine vitric ash lost above the vent and from the ash-flow later settles out as an extensive *co-ignimbrite* ash-fall.

Some eruptive-column collapse events generate *pyroclastic surges*, these being turbulent and highly expanded particulate systems that are outwardly directed or travel downslope as density flows. *Base surges* form when the eruptive column is overloaded with more or less cold pyroclasts, for example, juvenile magma chilled in contact with water (as in phreatomagmatic eruptions) or recycled rock debris (as in phreatic eruptions).

The classification of explosive eruptions by Escher (1933) correlated the style of activity with magma viscosity and gas pressure. One by Walker (1973) was a first attempt at quantification of eruptions by two parameters (D—dispersal index; F—fragmentation index) measured on the resulting pyroclastic deposits. One by Pyle (1989) depends on the fact that when the square root of the areas enclosed by isopleths is plotted against either deposit thickness or clast size, a straight line (exponential) relationship exists. Deposits can then be categorized by the derived quantities b_t (the distance over which the thickness halves), b_c (the distance over which the clast-size halves), and the half-distance ratio b_c/b_t .

GEORGE P. L. WALKER

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Cross-references: *Acid(ic) igneous rocks; Astrobleme; Calderas and volcano-tectonic depressions; Diatreme; Lava and lava eruptions; Nuée ardente; Pyroclastic eruptions; Pyroclastic flows; Rhyolite; Volcanoes and volcanism.*

EXSOLUTION—See Vol. IVA

F

FACIES OF THERMAL METAMORPHISM

Rocks in aureoles commonly show changes in mineralogy and texture (Harker, 1939) related primarily to distance from the contact with the intrusion. Within the broad spectrum of changes induced mainly by heat from the intrusion, four facies of thermal metamorphism are generally accepted. The four facies, listed in ascending order of probable maximum temperature attained, are albite-epidote hornfels, hornblende hornfels, pyroxene hornfels, and sanidinite.

Beginning about 1911, the concept of metamorphic facies was developed and extended by V. M. Goldschmidt, P. Eskola, and others (references in Turner, 1948). Eskola (1939) presented a detailed account of the facies principle and metamorphic facies that was widely accepted and remains the basis of most modern discussions. Eskola did not recognize separate groups of facies for thermal (contact) and regional (dynamothermal) metamorphism; his facies included pyroxene hornfels and sanidinite facies, while lower-grade hornfels assemblages were grouped with regionally metamorphosed rocks in the amphibolite and epidote amphibolite facies. Fyfe et al. (1958) divided metamorphic facies into two separate series for contact and regional metamorphism, and introduced the albite-epidote hornfels and hornblende hornfels facies in contact metamorphism. More recent formulations of the metamorphic facies series concept do not make such a sharp distinction.

Of the many factors influencing development of an aureole, two of the most important are temperature of the intruding magma and bulk composition and mineralogy of the host rock. Argillaceous and impure carbonate rocks, with initial assemblages of sedimentary or lowest grade metamorphic minerals, tend to develop the most distinctive contact-metamorphic parageneses.

Pitcher and Read (1963) related contact metamorphism to manner of emplacement of high-level granites in Donegal, Eire. They considered three main modes of emplacement: reactive, forceful, and permitted. In the first

case, intense reaction between the granitic magma and its envelope produced andalusite- and sillimanite-bearing hornfels. Strongly forceful intrusion developed contact schists with texture and mineralogy more akin to regional metamorphism; permitted intrusion caused only feeble contact effects. In most studies of aureoles, manner of emplacement is not explicitly considered as a control on development of metamorphic facies.

Albite-epidote hornfels facies

The albite-epidote hornfels facies develops in the outermost parts of aureoles. Commonly, the first obvious contact-metamorphic mineral is biotite, but in aluminous rocks andalusite or cordierite may appear first (Harker, 1939). The rocks rarely have the appearance of hornfels, but retain relict textures and structures that are partly modified by spotty development of new minerals. The spotty occurrence and incomplete development of mineral assemblages generally preclude mapping a continuous zone of rocks in the facies in an aureole. Mineral assemblages (Fig. 1) are similar to those in upper greenschist facies of regional metamorphism. The major differences are the occurrence of andalusite and absence of stilpnomelane and almandine-rich garnet in the albite-epidote hornfels facies

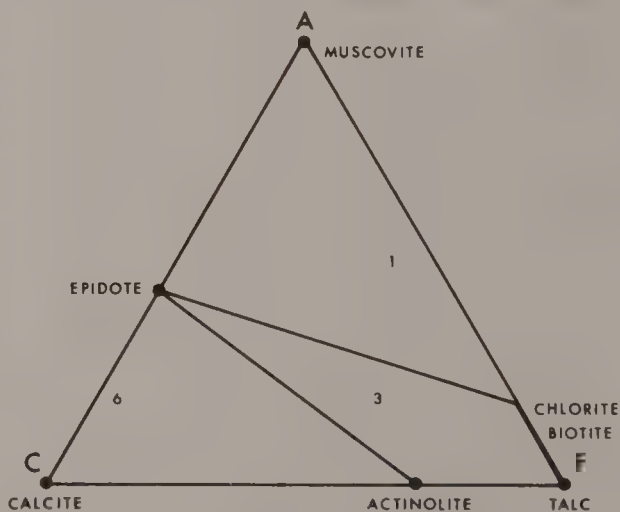


FIGURE 1. Albite-epidote hornfels facies; ACF diagram for rocks with excess SiO_2 and K_2O ; quartz, albite, and microcline are possible additional phases. (After Turner, 1981.)

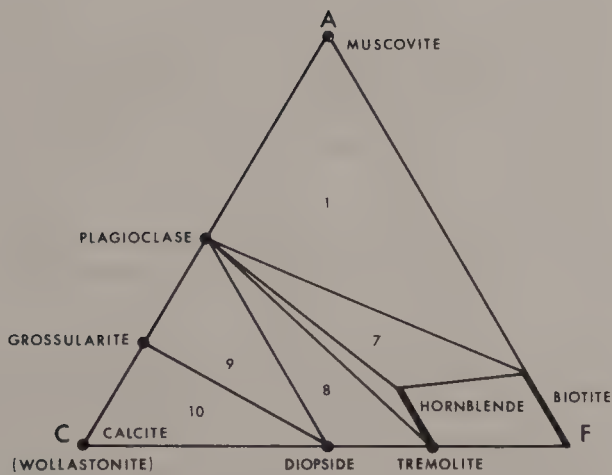


FIGURE 2. Hornblende hornfels facies; ACF diagram for rocks with excess SiO_2 and K_2O ; quartz and microcline are possible additional phases; numbers refer to assemblages observed at Orijärvi, Finland. (After Turner, 1981.)

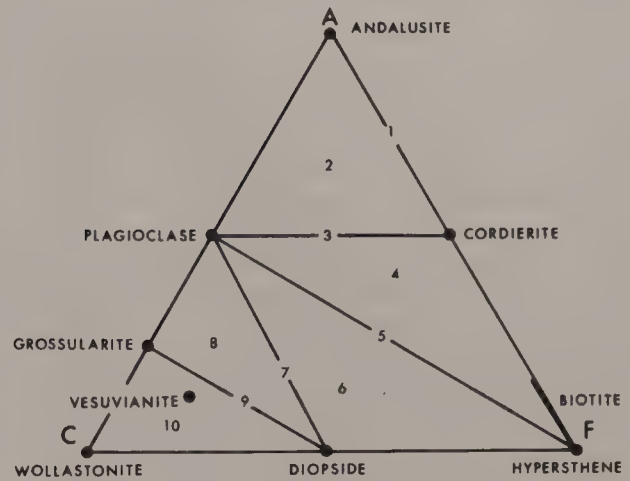


FIGURE 3. Pyroxene hornfels facies; ACF diagram for rocks with excess SiO_2 and K_2O ; quartz and K-feldspar are possible additional phases; numbers refer to assemblages observed near Oslo, Norway. (After Turner, 1981.)

(Turner, 1981). Albite is the expected plagioclase, although more Ca-rich varieties may persist as relics.

Hornblende hornfels facies

Assemblages of the hornblende hornfels facies are extensively developed in many aureoles. The facies may occur in the innermost aureole (and xenoliths) of granitic intrusions, but around intermediate and gabbroic intrusions may grade inward into pyroxene hornfels facies. Pelitic rocks in the hornblende hornfels facies typically have the dark-colored, massive appearance of hornfels, but in some localities retain a relict schistose or gneissose appearance. Andalusite and cordierite are present in most rocks of pelitic composition (Fig. 2). Common hornblende and plagioclase, generally more calcic than albite, occur in mafic rocks. Tremolite and diopside are developed in impure calcareous and dolomitic rocks. Staurolite, kyanite, and almandine are scarce, except in assemblages transitional with those of the amphibolite facies in regional metamorphism.

Pyroxene hornfels facies

The pyroxene hornfels facies (K-feldspar-cordierite hornfels facies of Winkler, 1967) is characteristic of innermost aureoles of gabbroic and ultramafic intrusions and of xenoliths in such intrusions. Most pelitic and many mafic rocks in this facies are hard, very dark-colored, and lacking in visible structure. Granoblastic textures are typical. The ACF diagram for this facies (Fig. 3) is from the classic interpretation by V. M. Goldschmidt (reference in Turner, 1948) of hornfels assemblages in the Oslo area,

Norway, and is the first ACF diagram for any facies. Andalusite occurs in lower-grade parts of the facies and sillimanite in the higher. Cordierite is a member of many assemblages. Hornblende and biotite tend to be dark brown to reddish-brown in color and are restricted to lower-grade parts. Augite and/or hypersthene are present in a wide range of compositions, hence the facies name. Muscovite and staurolite are absent. Assemblages containing one or more of the minerals diopside, grossularite, wollastonite, and forsterite are developed in calcareous and dolomitic rocks.

The progressive mineralogical changes by which assemblages of the pyroxene hornfels facies are formed are in most cases dehydration and decarbonation reactions (Winkler, 1967; Turner, 1981), and so the rocks are depleted in volatiles. Also, metasomatism involving removal of such elements as Si, K, Na, and Rb has been described from many localities, particularly in the upper range of the facies (see **Contact metasomatism**). Corundum-bearing hornfels are common in metasomatically altered zones.

Sanidinite facies

Assemblages of the sanidinite facies form in a restricted environment of very high temperature and low pressure, as narrow aureoles adjacent to shallow mafic volcanic necks, pipes, and dykes, and in xenoliths in volcanic and subvolcanic mafic rocks. Sanidine and pigeonite, which are restricted in metamorphic rocks to this facies, are present only because rapid cooling prevents inversion to lower-temperature polymorphs. Pyroxenes, cordierite, and calcic

plagioclase are typical minerals; micas and amphiboles are absent. The high-temperature polymorphs of silica (tridymite and cristobalite) may be present. Mullite and corundum occur in aluminous rocks. Silica-deficient carbonate rocks break down into assemblages containing such rare minerals as monticellite, melilite, spurrite, merwinite, larnite, and rankinite. The degree of metasomatism shows wide variations and apparently depends strongly on local conditions. Diffusion rates are high, but the time of metamorphism may be short. Temperatures may be intense enough to cause melting, particularly in pelitic rocks and impure sandstones. Rocks containing glass formed by melting during contact metamorphism are called *buchites*. Crystals of cordierite, corundum, mullite, or tridymite are common in the glass (Turner, 1981).

Conditions of metamorphism. Experimentally determined equilibrium curves which define stability limits for some individual minerals and simple assemblages important in contact metamorphism permit estimates of the conditions of metamorphism to be made—Fig. 4 shows estimated P - T fields for the four facies discussed here. The solid lines are dehydration and decarbonation curves; their positions would shift according to the relation of fluid pressure to total pressure and $P_{\text{H}_2\text{O}}$ to P_{CO_2} . In rocks containing significant proportions of the constituents quartz, plagioclase, and K-feldspar, melting would be expected to begin at or near the minimum melting curve when $P_{\text{H}_2\text{O}} = P_{\text{TOTAL}}$. Winkler (1967) interpreted the lower boundaries of the albite-epidote hornfels and hornblende hornfels facies to be 100–180°C higher than those in Fig. 4; his lower boundary of the pyroxene hornfels facies is approximately the same.

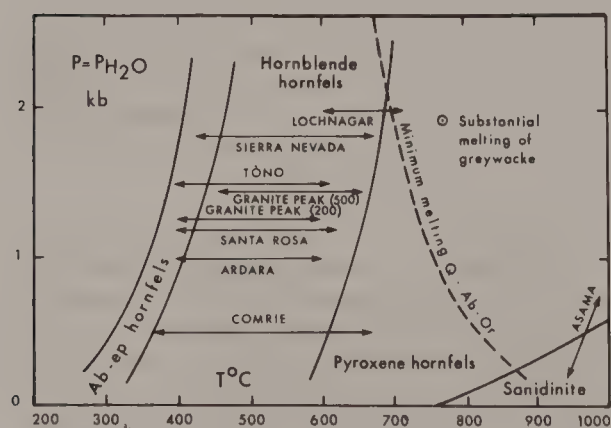


FIGURE 4. Diagram showing tentative P - T fields of the four facies of contact (thermal) metamorphism; inferred gradients, assuming uniform pressure, are shown for several aureoles. (After Turner, 1981.)

Other terms

Calc-flinta—a fine-grained calc-silicate rock of flinty appearance formed from calcareous mudstone, possibly with associated pneumatolytic action.

Calc-silicate rock—a rock composed mainly of Ca-bearing silicates (e.g., diopside, wollastonite) formed from impure limestone or dolomite.

Fleckschiefer—a spotted slate with minute flecks or spots of indeterminate material.

Fruchtschiefer—a spotted slate with concretionary spots having shapes suggestive of grains of wheat.

Garbenschiefer—a spotted slate with concretionary spots having shapes suggestive of caraway seed.

Knotenschiefer—a spotted slate with conspicuous spots that are often individual minerals (e.g., andalusite, cordierite).

Knotted hornfels facies—rock assemblage formed by low-grade contact (thermal) metamorphism—temperature 250–350°C, pressure < 2.5 kb; equivalent to albite-epidote hornfels facies).

Skleropelite—argillaceous-type rocks indurated by low-grade metamorphism without the development of cleavage (as in slate).

Spotted slate—an argillaceous slaty rock with a spotted appearance resulting from the incipient growth of porphyroblasts.

J. ROBERT BUTLER

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Cross-references: *Aureole*; *Batholith*; *Buchite*; *Contact (thermal) metamorphism*; *Contact metasomatism*; *Hornfels*; *Hornfelsic textures*; *Marble*; *Metamorphic facies series*; *Metasomatism*; *Pyrometamorphism*; *Pyrometasomatism*.

FELDSPATHOIDAL ROCKS—GENESIS

Feldspathoidal rocks are a special group of alkaline rocks mostly of igneous origin, that contain less silica than is necessary to form feldspar when combined with alkalis and alumina and therefore contain the feldspathoidal minerals, chiefly nepheline, leucite, and sodalite. They are volumetrically minor compared to silica-saturated and oversaturated rocks, and in both intrusive and extrusive environments are commonly associated with normal igneous types. They usually, but not always, contain high levels of the alkalis (Na_2O and K_2O) and are in many cases also peralkaline, i.e., they have molecular $(\text{Na}_2\text{O} + \text{K}_2\text{O}) > \text{Al}_2\text{O}_3$. This is shown by the presence of Na-bearing dark minerals, such as the pyroxenes aegirine-augite and aegirine, or the alkaline amphiboles, riebeckite and arfvedsonite. Some feldspathoidal rocks contain relatively high $\text{Na}_2\text{O}:\text{K}_2\text{O}$ (nephelinites, phonolites, and their plutonic equivalents—Fig. 1); others are distinctively rich in K_2O (called broadly the ultrapotassic rocks—exemplified by leucitite on Fig. 1). The problem for petrologists is to discover mechanisms by which uncharacteristically high levels of alkalis may develop by broadly “igneous” processes, and also how relatively silica-deficient rocks may evolve, bearing in mind that the continental crust is in general composed of oversaturated rocks (i.e., with excess silica). Magmas crystallizing feldspathoids have, in thermodynamic terms, low silica activity, and Carmichael et al. (1974) have classified igneous rocks on this basis.

Genetic schemes have to accommodate two subgroups of feldspathoidal rocks. On the one hand are feldspathoidal rocks that have compositions so exceptional that, for chemical reasons, they cannot be derived from the common magma types. Nephelinites and some ultrapotassic rocks fall in this category. On the other hand are rocks such as phonolites, which may be derived by crystal fractionation processes from more common parents, such as trachytes and alkaline basalts (Fig. 1). Although modern theories, in the light of overwhelming chemical and isotopic evidence, place the ultimate source of most feldspathoidal rocks in the upper mantle, there is also clear evidence, from the associations of rocks found in

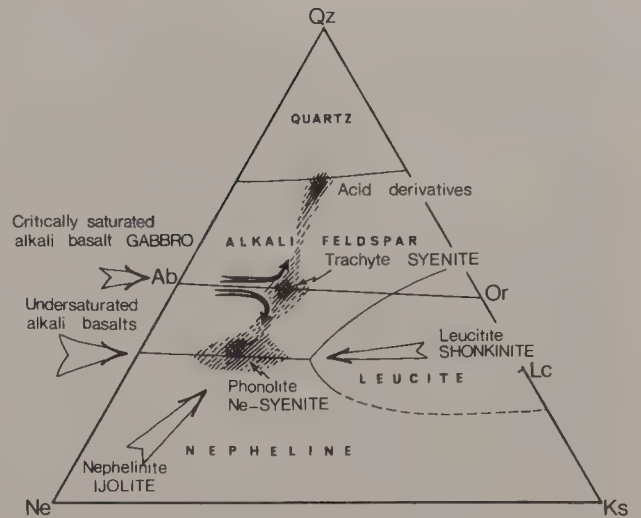


FIGURE 1. Generalized diagram of “petrogeny’s residua system,” nepheline (Ne)–kalsilite (Ks)–quartz (Qz); primary phase areas labeled in bold; large arrows illustrate lineages that lead into the system from the more calcic and, usually, more Mg–Fe-rich parents. Important rock types that approximate to the system are shown—extrusive rocks in lower case, intrusive equivalents in capitals. The shaded areas show the concentrations of rock analyses (as normative constituents) observed in nature; these lie on low-pressure thermal minima or in “thermal valleys” in both silica-oversaturated and undersaturated parts of the system, and the syenites plot on the thermal minimum on the feldspar join Ab (albite)–Or (orthoclase), which is a thermal ridge.

individual volcanoes or plutonic igneous complexes, that low-pressure crystal fractionation has contributed to the formation of the spectrum of feldspathoidal rock types. To investigate the ultimate source of the undersaturated magmas, it is necessary to look back through this evidence of relatively high-level magmatic evolution in order to deduce the probable nature of the least evolved magmas that emerged from the mantle.

Certain very characteristic features of the occurrence of feldspathoidal rocks also have to be accommodated in petrogenetic schemes. Although alkaline types do occur in oceanic islands, by far the majority of examples come from stable anorogenic continental environments and are usually associated with rift valley formation. Nephelinite volcanism is frequently explosive, and carbonatites are very commonly present. Le Bas (1977) has provided the most complete description of this type of magmatism, in the setting of the present-day E African rift valley, and a superb example of a Proterozoic rift with a carbonatite–nepheline syenite association is afforded by the Gardar province in S Greenland (Emeleus and Upton, 1976). Metasomatism (the modification of wall rocks by

alkali-charged hydrothermal fluids) is widespread in feldspathoidal intrusions, and has led to many controversies. Bailey (1977) has drawn attention to the long-lived nature of alkaline magmatism in the African system (ca. 750 Ma to Recent), and in the much older Gardar rift feldspathoidal rocks were generated over a span of >150 Ma, very much longer than, for example, the time taken for the North Atlantic to open. Another significant observation is that adjacent volcanoes may respectively produce highly sodic and highly potassic lavas. As well as enrichment in alkalis, extreme feldspathoidal rocks are also highly enriched in unusual elements, such as Zr, Be, Nb, and rare-earth elements (REE). The extraordinary naujaite (eudialyte sodalite syenites) of Ilimaussaq, S Greenland, contain 5 wt.% ZrO_2 and 0.00 wt.% MgO, are truly the ultimate products of the Earth's igneous fractionation.

A review of feldspathoidal rocks is given by Fitton and Upton (1987).

Fractional crystallization

The mechanism of fractional crystallization must be dealt with first because there is ample evidence that many bodies of nepheline syenite and associated syenitic rocks have mineralogies related by crystal fractionation processes, as they can be deduced from experimental studies at comparatively low pressures. By exploring such lines of descent it is possible to locate potential barriers to fractionation and then look elsewhere for the ultimate source of the magmas.

Some of the problems of alkaline rock genesis can be best understood by reference to Fig. 1, which illustrates experimentally determined melting relations in "petrogeny's residua system," and also serves to summarize some of the more important rock names (see papers by MacDonald and Edgar in Sørensen, 1974). Syenites and nepheline syenites have normative mineralogies very close to thermal minima in this system. Various lineages lead to these compositions. The most important is that starting with just-saturated or undersaturated alkali basalts:

alkali basalt–hawaiite–mugearite–
benmoreite–trachyte–phonolite.

This sequence appears commonly in both oceanic and continental volcanic suites, and in the plutonic setting in the derivation from gabbro of syenite and nepheline syenite. Other, more undersaturated lineages exist, e.g.,

basanite–nepheline hawaiite–phonolite,

and potassic lineages

leucite basanite–leucite.

The crystallization paths followed depend essentially on the initial magma composition. Basalts close to the critical plane of silica-undersaturation with neither appreciable hypersilica nor nepheline in the norm tread a delicate path that leads to trachyte. The importance of the "thermal barrier" along the feldspar join (Fig. 1) is that, on strong fractionation, trachytes (or syenites) slightly to one or the other side of this barrier will have their initial slight degree of over or undersaturation accentuated by fractionation, to lead eventually to the granite and nepheline syenite minima.

Until perhaps the last 20 years, petrologists sought to generate all basaltic lineages from one or two truly primitive magma types, but the modern view is that a spectrum of basaltic to nephelinitic liquids, with a range of silica activity, can be produced by melting at different depths in the mantle. These liquids may evolve at various stages during their uprise through the upper mantle and crust, and the fact that their chemistry is so closely related to low-pressure equilibria in the residua system (Fig. 1) shows that their evolution has continued to a high level. A very few intrusive complexes have compositions that lie in both over- and undersaturated fields, perhaps the best known being the Kangerdlugssuaq complex in E Greenland (Pankhurst et al., 1976), but the majority of complexes have rock compositions that plot in one or another syenite field. Modification of an undersaturated syenite by silica-charged hydrothermal fluids was suggested to explain the case of Kangerdlugssuaq. Normally, crystallization of alkali feldspar from an even mildly undersaturated liquid must drive the magma towards the "nepheline syenite" minimum.

Origin of alkali basalt, nephelinite, and leucite-bearing assemblages

Nepheline–pyroxene rocks (nephelinite, with ijolite the intrusive equivalent) occur in many major alkaline provinces, often together with carbonatites. Normal schemes of fractionation of alkali basalts cannot lead to nephelinite, and it can be seen in Fig. 1 that in this simplified system ijolite cannot be reached from syenitic parents because of the thermal valley occupied by the nepheline syenites. However, Bailey and Schairer (1966) have shown that when the melting relations of more complex peralkaline starting materials (i.e., acmite normative) are considered, ijolite can be reached by slight partial melting of slightly undersaturated per-

alkaline syenite. However, this mechanism cannot explain the association of nephelinite with nepheline basalt, melilite basalt, and melilitite, or the outpouring of huge volumes of nephelinite in the E African rift.

There is evidence, from included peridotite and eclogite nodules, that nephelinites come from great depths. Chemically, they have close similarities to the kimberlites, which also have a high-pressure origin. Green (1970) reviewed experimental evidence that partial melting of hydrous pyrolite (a probable mineral assemblage for the upper mantle) could lead to production of very undersaturated basanite and olivine nephelinite liquids, the former in equilibrium with magnesian olivine, orthopyroxene, and calcic clinopyroxene in the 18–35 kb range of pressure and the latter in the 18–22 kb range. The presence of garnet in the residual crystalline phases will produce early partial melts with low Al_2O_3 and higher CaO contents, giving rise to olivine–melilite nephelinites. At $P < 18$ kb, amphibole will appear at the solidus at high $P_{\text{H}_2\text{O}}$. Liquids in equilibrium with amphibole at low degrees of partial melting are probably relatively K-rich, ranging from basanites at 18 kb to olivine trachybasalts at 9 kb. The importance of this type of partial melting mechanism for the derivation of alkaline liquids lies in the flexibility it provides as a source for a spectrum of basaltic types from nephelinite to oversaturated tholeiites, dependent upon depth of melting and water content. It appears to explain much of the diversity observed in alkaline volcanicity.

Several modern investigations of feldspathoidal rocks have suggested that to seek an origin in “normal” mantle is incorrect. Petrographic investigations of mantle xenoliths in alkaline rocks (Lloyd, 1981) have suggested that the mantle was modified by metasomatic events prior to separation of the alkaline melts. This “enrichment event” not only provides a source of alkalies, but also transports to the source region trace and minor elements characteristic of alkaline rocks. Bailey (1977) pointed out that the longevity of alkaline magmatism in many regions implied that the mantle under such regions was modified in a characteristic way, and that once modified it could repeatedly produce alkaline magmas. He suggested that updoming of a lithospheric plate (as seen in E Africa) by forces external to the plate caused a low-pressure zone in the upper mantle to which CO_2 -rich, alkali-laden fluids migrated. Ultimately, this may have led to the present nephelinite–carbonatite volcanism on the floor of the rift system, and also accounts for the older nepheline syenite plutonism.

Leucite-bearing rocks, which have the highest

known K_2O contents for silicate rocks, are of very limited occurrence and pose special problems of petrogenesis, both because they cannot be reached by fractionation of more common rock types (Fig. 1) but also because K_2O occurs at very low levels in mantle peridotites. This means the melt must be separated from a relatively large volume to produce liquids of such high K_2O content. The further fractionation of such leucite-rich melts into the nepheline syenite minimum (“shonkinite” trend—Fig. 1) has been described for the Shonkin Sag laccolith by Nash and Wilkinson (1970). A source of K_2O in mantle rocks that has been invoked is phlogopite mica, which is stable at upper-mantle pressures. Modification of mantle by subduction of crust has been postulated as one source of phlogopite in the celebrated Roman ultrapotassic region, and complex, multistage processes of mantle enrichment have been invoked to account for the trace element and isotope systematics of this region (Rogers et al., 1985). However, in other provinces, the coexistence of ultrapotassic and strongly sodic magmas, and their simultaneous outpouring from adjacent volcanoes, would suggest that much more localized processes can lead to this dichotomy.

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Cross-references: *Alkaline rocks—undersaturated; Basalt; Carbonatites; Leucitite; Magmatic differentiation; Metasomatism; Nephelinite; Peralkaline igneous rocks; Phonolite; Rift valley volcanism; Syenite.*

FENITE

The term *fenite* was proposed by Brögger (1921) for a group of metasomatically altered rocks in and adjacent to the alkaline complex at Fen, Norway. The metasomatizing solutions were presumed to have emanated from an ijolite–melteigite magma reservoir and resulted in the development of rocks of alkali-syenitic composition from rocks originally of granitic composition. The associated *tveitåsites* were regarded as contaminated rocks but subsequently von Eckerman (1948) showed that comparable rocks on Alnö Island, Sweden, resulted from metasomatism.

The term *fenitization* has, in time, acquired a wider meaning than Brögger gave to it, and is now in common use for processes of metasomatism whose effects are seen in the immediate vicinity of carbonatite complexes and around bodies of nephelinite magma that fractionated and crystallized to ijolite. The effects of fenitization are more varied around carbonatites than around ijolites but with sövites (rarely microsövites) and dolomitic carbonatites rather than alvikites and ferrocarbonatites affected. This is taken to indicate that alkalis are lost during early stages of carbonatite fractionation. The intensity of the metasomatism commonly causes the original rock composition to be obscured totally (Fig. 1).

Mineralogical and chemical changes that take place are illustrated by a suite of fenite nodules in the Oldoinyo Lengai carbonatitic volcano located in the southern part of the Gregory Rift, N Tanzania where temperatures of ca. 800°C were attained during metasomatism (Morogan and Martin, 1985). Fenites with a granitic protolith developed in three stages with relict

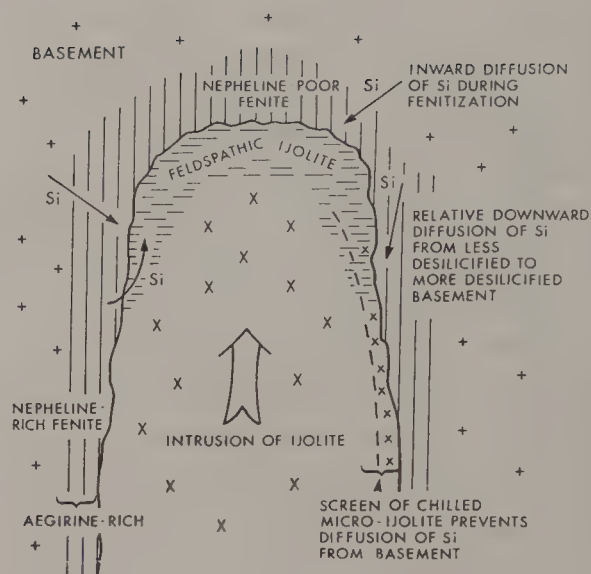


FIGURE 1. Schematic cross-section of an intrusion of ijolite with the development of fenite in its marginal parts and in the adjacent host rock. (After Le Bas, 1987.)

minerals (oligoclase, biotite, quartz) + aegirine + anorthoclase → anorthoclase + aegirine-augite → K-sanidine + aegirine-augite + calcite ± nepheline ± hematite. These changes are reflected in strong depletion of SiO₂ in the protolith with addition of alkalis, CaO, Fe₂O₃, and possibly TiO₂. The Na₂O/K₂O ratio of low- to medium-grade fenites is high, but decreases markedly in the highest grade fenites.

Factors involved in fenitization include (1) chemical composition of the magma causing the metasomatism, (2) the nature of the metasomatizing fluids, (3) the mineralogical composition, chemical composition, porosity, and structure of the rocks being fenitized, (4) the depth, and (5) any water–rock reaction (cf. Woolley, 1982).

The complexity of the fenitization processes and the diversity of its products is reflected in the nomenclature of fenites. Terms such as quartz syenite fenite and nepheline syenite fenite (e.g., von Eckerman, 1948) emphasize the nature of the igneous intrusion, while such terms as fenitized granite and fenitized gabbro emphasize the nature of the protolith (e.g., Robins, 1984). Terms such as sodic fenite, potassic fenite, and intermediate fenite reflect the importance of alkalis and the Na/K ratio, while spatial relationships and varying grade of fenitization are reflected in terms such as low-, medium-, and high-grade fenites, and contact, aureole, and vein fenites.

Much study of fenite has taken place in eastern and southern Africa with early records of production of almost pure K-feldspar rock around the upper parts of masses of sövite coming from Malawi (Dixey et al., 1935),

Tanzania and Zambia (cf., Le Bas 1977, 1987). The K-feldspar in these rocks is usually orthoclase (Or_{87-95} ; the rocks are termed *orthoclasites*), but in feldspathized ijolite and ijolitic fenite it is less potassic (Le Bas, 1984). Phlogopite-rich rather than feldspar-rich fenites are formed where sövite intrudes mafic rocks (Gittins et al., 1975). In more deeply eroded and older complexes albite-rich fenites (*albitites*), sometimes containing Mg-arfvedsonite, aegirine, and Fe-oxides, are found around sövites, for example, on the Kola Peninsula and between the Azor and Caspian seas, U.S.S.R. (Kapustin, 1981) and in W Tanzania (Coetzee, 1963). Albitite also occurs as blocks brought up by carbonatitic volcanic breccias in complexes in which K-rich fenites are exposed at the surface, while the principal carbonatite complexes of Malawi have outer sodic fenite aureoles that are replaced by minor potassic fenite aureoles (cf., Woolley, 1982).

The mineralogy and mineral chemistry of the Fen and Alnö complexes are given by Kresten and Morogan (1986) and Morogan and Woolley (1988), respectively, together with discussion of genesis and extensive bibliographies. Reviews of African and other examples of fenites, and extensive bibliographies, are given by McKie (1966) and Le Bas (1977, 1987).

DUNCAN MCKIE

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FISSION TRACK DATING—See Vol. IVA

FLOOD BASALT

Flood basalts form thick piles of basaltic lava flows of vast areal extent (often exceeding 100,000 km²) and volume (exceeding 100,000 km³), usually on continental crust. Individual flows may be very voluminous, both exceeding 20 m in thickness, and covering more than 1,000 km² (Table 1). Individual lava plateaus are usually associated with within-plate hot spots or intraplate rifting (or both) and have a number of features in common.

1. Early stages of eruption history are generally associated with regional dyke swarms, some of which fed fissure eruptions.
2. Later stages of activity may involve the establishment of central volcanic complexes, usually shield volcanoes.
3. Early fissure eruptions produce both tholeiitic and alkaline basalts.
4. Shield volcanoes often contain more strongly differentiated lavas and intrusive rocks.

Flood basalts with these characteristics have developed since late Archaean times, on all the continents, and in the case of Iceland, on oceanic crust as well. Examples include the early Proterozoic Jatulian lavas of E Finland and Soviet Karelia, the mid-Proterozoic Keweenawan lavas and Coppermine River lavas of N America, the Carboniferous Clyde Plateau

TABLE 1. Data on flood basalt provinces

	Present area (former area)(km ²)	Maximum thickness (m)	Approx. volume (km ³)	Age	Duration of activity (Ma)	Reference
Keweenaw (U.S.A.-Canada)	100,000	17,000	1,000,000	Mid-Proterozoic	100	Baragar (1977)
Siberian Platform (U.S.S.R.)	337,500 (700,000)	4,000	337,500	Permian-Triassic	50	Masaitis (1969)
Karoo (S Africa etc.)	140,000 (2,000,000)	2,000	140,000	Jurassic-Cretaceous	50	Cox (1972)
Paraná (Brazil etc.)	1,200,000	2,000	780,000	Jurassic-Cretaceous	30	Bellieni et al. (1984)
Deccan (India)	600,000 (1,000,000)	2,000	780,000	Cretaceous	5	Bose (1972)
Colombia River (U.S.A.)	220,000	4,000	200,000	Tertiary	9	Wright et al. (1973)
Snake River (U.S.A.)	200,000	4,000	200,000	Tertiary	5	Wright et al. (1973)
Onega basin (U.S.S.R.)	ca. 200,000	5,000	ca. 200,000	Early Proterozoic	100	Heiskanen (1980)
British Tertiary	7,000 (10,000)	2,000	ca. 100,000	Tertiary	5	Emeleus (1983)
W Greenland	55,000	3,000	ca. 110,000	Tertiary	5	Clarke and Pedersen (1976)
E Greenland	40,000	2,500	66,000	Tertiary	5	Deer (1976), Noe-Nygaard (1976)



FIGURE 1. Distribution of some major flood basalt provinces; see Table 1.

lavas of central Scotland, the Jurassic and Cretaceous Gondwanan lava plateaus of S America, S Africa, Antarctica, Malagasy and India, the Triassic lavas of the Siberian Platform, and the Tertiary lava plateau of northwestern Greenland and the British Isles (Fig. 1; Table 1). Similar massive eruptions may also characterize other terrestrial planets, such as the lunar maria and the Tharsis Ridge on Mars.

That flood basalts form plateaus of varying size, with little interflow material, is reflected by the often-used term trap or trappe (from Swedish, "trappa", stairs or step-like). Most flows were erupted on land, so that palaeosols are usually the only interflow material. Where lakes were impounded and then engulfed by lava, hyaloclastic rocks were produced; where crater-lakes occupied central calderas, pillow lavas may have developed. Eruptions often occurred over a landscape with very little relief, though the huge thickness of lavas accumulated may have produced crustal downsagging. Most lava plateaus appear to have accumulated very rapidly.

Where flood basalts are related to continental rifting and break-up, major crustal structures control the disposition of the regional feeder-dike swarms, e.g., the E Greenland coastal flexure, and the Lebombo monocline in S Africa (Fig. 2). The deep-levels of rift zones may be

the sites for the emplacement of large stratiform basic intrusions, e.g., the Noril'sk Talnakh, (Siberia, U.S.S.R.), Muskox (Canada), and Duluth (U.S.A.) intrusions. Subvolcanic complexes including ring intrusions are also a conspicuous feature of these plateaus, and they may include both layered gabbros (E Greenland; NW Britain), granites (NW Britain; Victoria, Australia) or strongly differentiated alkaline or peralkaline intrusions (Gardar, S Greenland; Keweenaw, U.S.A.-Canada; Kangerdlugssuaq, E Greenland).



FIGURE 2. Flood basalts and continental break-up: (a) Jurassic-Cretaceous flood basalts of the Gondwana supercontinent and (b) Tertiary flood basalt provinces of the N Atlantic region.



FIGURE 2. (continued)

Petrography

The vast bulk of flood basalt plateaus consists of monotonous accumulations of both tholeiitic (quartz- and hypersthene-normative) and alkaline (nepheline-normative) basalts. A number of suites of basalts and related rocks are evident; these include tholeiitic basalt-alkali-rich rhyolite, alkaline basalt-mugearite-hawaiite-trachyte or phonolite, and alkaline basalt-basanite. More than one such trend may be evident on one plateau, and such trends may show a cyclic development with time, or one trend with time may dominate a whole province. The more monotonous basalt accumulations are generally related to fissure swarms, and the dykes show the same uniformity of composition; either olivine or quartz dolerites, with picrites and foidal varieties like teschenite occur. Localized dyke swarms within the plateaus can include quite a variety of alkaline mafic, highly potassic and ultramafic rock types, so that lamprophyres like monchiquite-camptonite, melilitic rocks, nephelinites, and limburgites are quite widespread, especially where intracontinental rifts are present. In the Gondwanan basalt plateaus (particularly S Africa—Fig. 2a) the Jurassic and Cretaceous lavas have a secular and spatial relationship to kimberlite pipes, and in Siberia there are kimberlitic lavas (meymechites).

Like kimberlites, nephelinites, and lamprophyres in these provinces, the basalts frequently carry inclusions, both cognate and accidental. These include ultramafic peridotite, eclogite, and granulite, with mineralogy indicating a deep crustal or upper mantle origin.

Chemical composition

Table 2 presents major- and trace-element analyses of averaged representative basalts from four flood basalt plateaus. The major element compositions reflect the spectrum of hypersthene- to nepheline-normative and feldspathic (trachy- and andesitic basalts) to picritic types. All are low in K_2O (<2.0 wt%) and display a wide range in TiO_2 (0.5–3.6%). Trace-element concentrations show similar quite wide ranges. In terms of overall composition, this reflects the gross similarity of basalts from all their various tectonic settings.

Incompatible element patterns, on the other hand, display many characteristics that are unique to flood basalts. These characteristics are picked out on the variation diagrams referred to as “spidergrams,” normalized to mid-ocean ridge basalts (MORB). On these diagrams, the island-arc basalts (IAB), ocean-island basalts (OIB), and continental flood basalts (CFB) show many similarities (the characteristic downward-concave curvature for instance) all distinctly different from MORB. But CFB generally show strong enrichment over MORB for Sr, K, Rb, Ba, Th, Ta, Nb, Ce, and P, with less for Zr, Hf, light rare-earth elements (LREE) and Ti, while Y and heavy REE show relative depletion. A similar pattern is shown by OIB, but there is less-marked enrichment in the group from Sr to Ce (Fig. 3). This is in striking contrast to the IAB.

Isotopic data are likewise fairly distinctive. The initial $^{87}Sr/^{86}Sr$ ratio ranges from 0.703 to <0.706, while initial $^{143}Nd/^{144}Nd$ ranges from 0.5130 to <0.5124; these are respectively higher and lower than MORB. Significantly, they overlap the model values for the bulk Earth— $^{87}Sr/^{86}Sr$, 0.7046, $^{143}Nd/^{144}Nd$, 0.51262).

Petrogenesis

The isotope, trace-element, and major-element analyses point to flood basalts having crystallized at liquidus temperatures, and having equilibrated with the mantle compositions inferred from studies of cognate xenoliths. At the same time, this mantle appears to have been less depleted than the MORB source region, perhaps enriched by upwelling material from the deep mantle. Variations are due to (1) the assimilation into and contamination of basaltic liquids by crustal material, and (2) the melt having been derived from regions of the upper mantle in which there are both secular and spatial compositional heterogeneities that reflect differing degrees of replenishment from the chondritic reservoir and such localized processes as K-metasomatism.

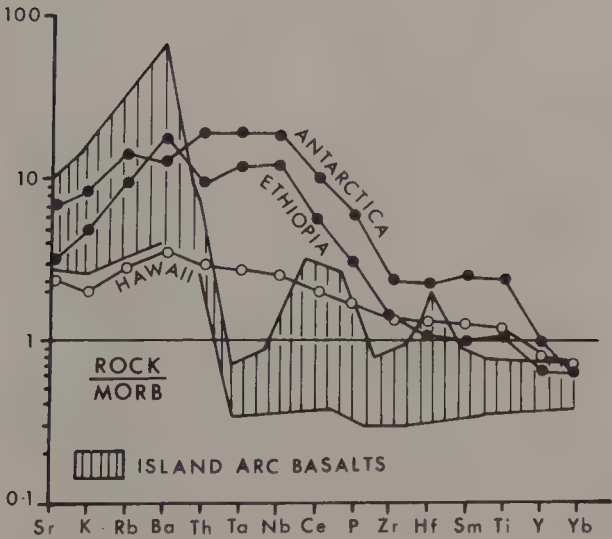
ADRIAN F. PARK

TABLE 2. Major- and trace-element analyses of typical flood basalts

	Columbia River	Antarctica	E and W Greenland	Ethiopia
SiO ₂	50.33	54.46	46.55	46.39
TiO ₂	2.44	0.61	2.23	2.09
Al ₂ O ₃	14.78	14.90	12.03	16.29
Fe ₂ O ₃ ^a	14.09	9.93	14.28	12.77
MnO	0.20	0.16	0.19	0.19
MgO	6.93	7.02	12.65	7.53
CaO	9.04	10.85	9.62	10.22
Na ₂ O	3.05	1.79	2.04	4.02
K ₂ O	1.11	0.90	0.57	0.96
P ₂ O ₅	0.68	0.08	0.17	0.43
Ba	644	156	247	325
Be	1.6	0.6	0.23	1.05
Cr	211	102	407	127
Cu	52	52	103	66
Ga	20	16	18	19
Hf	8.27	2.48	3.85	4.04
Nb	36.4	3.7	12.0	36
Ni	83	66	421	79
Pb	8	11	3	4
Rb	23	34	12.4	20
Sr	346	126	250	554
Ta	2.03	—	1.28	2.60
Th	2.89	3.58	1.26	2.56
W	—	—	41	—
Y	55	22	27	26
Zn	107	70	97	36
Zr	394	81	140	180
La	22.5	8	14.6	21.8
Ce	50.5	22	42.6	50
Pr	5.5	2.7	5.2	5.3
Nd	26.6	11	19.2	25
Sm	6.3	3.0	4.9	5.3
Eu	3.06	0.80	1.66	1.76
Gd	6.4	3.0	5.15	4.9
Dy	5.7	3.41	4.73	4.32
Ho	1.7	0.7	0.9	0.82
Er	2.7	2.15	2.5	2.25
Yb	2.9	2.19	2.15	1.94
Lu	0.44	0.33	0.31	0.28

Source: averages from data in Thompson et al. (1983).

^a Total Fe as Fe₂O₃.



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FIGURE 3. Incompatible elements from flood basalts (Antarctica and Ethiopia) plotted in a MORB-normalized spidergram for comparison with MORB, OIB (Hawaii), and IAB (Tonga, Fiji, Scotia arc, Kermadec, and New Hebrides). (After Pearce, 1983.)

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Cross-references: *Basalt*; *British Tertiary Volcanic Province*; *Igneous rocks—tectonic setting*; *Lava and lava eruptions*; *Mafic and ultramafic inclusions in igneous rocks*; *Rift valley volcanism*; *Ultramafic lavas*; *Upper mantle—experimental petrology*.

FLUID INCLUSIONS—See Vol. IVA

FLUIDIZATION

Fluidization is an industrial process in which gas is passed through a bed of fine-grained solid particles in order to facilitate chemical reaction between gas and solid. Geological illustrations of the process are provided by the infillings of some volcanic pipes and dykes, and the process has theoretical applications in explaining the high degree of mobility of nu es ardentes, the

rounding of blocks within volcanic vents, and the changed composition of fragmental rocks within vents (Reynolds, 1954, 1969; Zeng and Othmer, 1960).

At a particular rate of gas flow, appropriate to the size and density of the particles concerned, particles become separated from one another and freely suspended within the rising gas streams. As a result, the initial static bed of particles expands upward (Fig. 1a). Because the expanded bed (Fig. 1b) resembles a liquid in its physical properties—i.e., it has a horizontal upper surface, and measurable viscosity and density—the solid particles within it are said to be *fluidized*. Within a fluidized bed, large bodies with a relatively high density will sink, and those with a relatively low density will rise.

When the rate of gas flow is increased, the bed expands still further until a stage is reached when it presents the appearance of a boiling liquid. This is because, in addition to gas flowing uniformly through the expanded bed, gas now also rises as bubbles that disturb the surface of the bed. The bubbles differ from gas bubbles in a liquid, however, because there is a continual interchange between them and the gas of the expanded bed. The bubbles cause violent agitation of the fluidized bed, which under these conditions is known as the *boiling bed* (Fig. 1c); the particles within it show turbulent movements and in consequence become thoroughly mixed. Moreover, because the particles serve as heat carriers and make intimate contact with the gas, an even distribution of temperature is maintained throughout the boiling bed. As a result, the boiling bed provides ideal conditions for chemical reaction between gas and solid particles.

Fluidized systems erode the walls of their containing vessels and the particles within fluidized systems become abraded and rounded—the form that is most resistant to erosion being spherical.

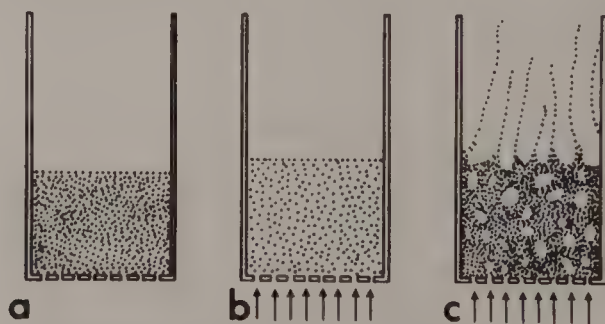


FIGURE 1. Stages in the process of fluidization: (a) initial static bed of fine-grained solid particles; (b) expanded bed—gas flows through perforations in the base of the containing vessel in the direction of the arrows; (c) boiling bed—gas bubbles rise upwards through the expanded bed, above which the finer-grained particles are entrained by the gas.

Geological examples

Volcanic vents. Within some volcanic vents, the larger blocks retain their initial orientation relative to one another, and to the wall-rocks from which they are derived, although they may have subsided or risen relative to their stratigraphic position. That the tuff-matrix, in such examples, was fluidized at the time when the blocks subsided or rose is evidenced by the manner in which it has penetrated structural breaks in the country rocks, eroding and widening them until blocks have become detached, stoped, and engulfed within the activated tuff (see **Tuffisite**). Yet structures like the dip and strike of sedimentary rocks, and the strike of detached portions of dykes that intersect them, indicate that the orientation of the detached blocks is undisturbed. Within the fluidized tuff, such detached blocks sank or rose accordingly as their density was greater or less than that of the fluidized tuff.

Examples are provided by the Swabian tuff-pipes, W Germany (Cloos, 1941, in Reynolds, 1954; and see **Diatreme**), and by some vents in the British Tertiary Volcanic Province, for example, near Kilchoan, Ardnarmurchan, NW Scotland (Reynolds, 1954), the Carboniferous vents of Fife, E Scotland (Francis, 1960), and the intrusive tuffs of western Cork, Eire (Coe, 1966).

Nuées ardentes. The fluid-like behavior of ash flows, in which an assemblage of particles of pumice and ash pour down volcanic slopes at high velocities, suggests that they travel as fluidized systems. Calculations by Brown (1962) and McTaggart (1962) indicate that the volume and rate of emission of volatiles (steam at 700°C was used for the calculations) from pumice and ash particles in nuées ardentes are amply sufficient to fluidize the assemblage of particles in the time required for their emplacement. An additional or alternative method of fluidization may be provided by escaping air that has been entrapped by the advancing flows.

Druses. An example corresponding to the bubble-phase of industrial fluidized systems is that of druses in a vein that intrudes upwards from a Triassic conglomerate into an overlying dolerite sill, near New Haven, Connecticut, U.S.A. (Walton and O'Sullivan, 1950, in Reynolds, 1954). Attached to the walls of the druses are crystals of quartz, epidote, pyrite, garnet, and sphene, together with nodules of chlorite. Their development was the result of vaporization of water within the interstices of the conglomerate due to the heat of the sill. In response to fall of pressure caused by fracturing of the cooling sill, the water vapor expanded upwards into the fracture with sufficient speed

to expand the conglomerate and so lift it upwards into the fracture. The mineral contents of the druses indicate that Si, Al, Fe, Mg, and Ti diffused through the water vapor.

Rounded boulders. Rounded boulders and pebble-like vent infillings in Eire (Reynolds, 1954) and Arizona, U.S.A. (Barrington and Kerr, 1961) are interpreted as examples of abrasion and rounding resulting from fluidization, while the constant proportions of the various grain sizes throughout the 300 m of known depth in a breccia pipe in Colorado, U.S.A. (Peters, 1953, in Reynolds, 1954) is considered to result from the action of the same process (see **Breccia pipe**).

DORIS L. REYNOLDS

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Cross-references: *Breccia pipe*; *Cryptovolcanic structure*; *Diatreme*; *Nuée ardente*; *Tuffisite*.

FUMAROLE

A *fumarole* is a vent issuing wet or superheated steam contaminated with various gaseous compounds, mainly volcanic gases such as CO₂, H₂S, and H₂. Fumaroles occur in all major areas of volcanic activity and four main types of fumarolic activity are recognized: (1) a high-temperature fumarole, generally of the solfataric type, issuing mainly superheated steam with a relatively high content of volcanic gases; (2) a steam hole, issuing saturated or wet steam; (3) a mud crater, which is a fumarole issuing from clayish overburden; and (4) hot ground, where steam condenses at the surface.

In most cases, the high-temperature fumarole is a postvolcanic phenomenon located at the

central crater, or on the flanks, of volcanoes that have been active in very recent times. Such fumaroles may also be found on the surface of newly-formed lava beds. The gases expelled by these vents have temperatures ranging from slight superheating up to several hundred degrees Celsius; the highest figure reported is ca. 600°C.

The steam hole is the most common type of steam outlet in high-temperature thermal areas and the term *soffione* is used for steam-type fumaroles although the term was originally applied to the boric acid fumaroles of Tuscany, Italy. Saturated steam issues from holes in the ground with diameters of a few centimeters up to several meters. The steam flow varies greatly; small holes may issue only a few tens of kilograms per hour of steam, whereas the greatest steam holes have a steady flow of the order of 10 tonnes per hour. The velocity of the escaping steam varies from <10 to >100 m/sec. Considerable noise is associated with the higher velocities. The rock surrounding the vents is subjected to a considerable thermal alteration due to high temperatures and the corrosive action of the volcanic gases intermixed with the steam.

The mud crater is a common type of fumarole that evolves in places where there is a moderate upflow of natural steam into surface clay or mud. In general, the clay is the result of a thermal weathering of the country rock. The steam penetrates the clay cover in many locations and condensed water helps to soften and emulsify the clay around the small outlets. Steam bubbles are formed in the mud and these explode upon reaching the surface, throwing mud and clay around. The result is a small crater-formed heap around the vent. The size of mud craters depends on the steam supply, and the thickness and consistency of the clay. Small craters may have a diameter of <1 m and the most common types have diameters of the order of a few meters. Exceptionally, great mud craters are found in the thermal areas of New Zealand where the surface is covered by a thick layer of light pumice. Some of these craters have diameters of several tens of meters. The temperature of mud craters is at, or close to, the boiling point of water, which depends somewhat on the altitude.

A fumarolic vent in a lava flow formed by a gaseous explosion in lava that is still fluid (probably owing to the generation of steam from underlying wet material) is called a *spiracle*. These are generally ca. 1 m across and up to 5 m high, although in the western U.S.A.

TABLE 1. Composition of gas in fumaroles

	Volume (%)		
	1	2	3
CO ₂	81.3	98.8	86.3
H ₂ S	12.5	0.1	2.0
H ₂	4.5	0	9.8
CH ₄	0	0.1	0
N ₂ + Ar	1.7	1.0	1.9

- 1. Steam hole, Krysuvik thermal area, Iceland.
- 2. Mud Volcano group, Yellowstone National Park, U.S.A.
- 3. Postvolcanic fumarole, White Island, New Zealand.

they are up to 10 m across and 12 m or more high.

The water issued by fumaroles may be of meteoric or juvenile origin, with isotopic evidence indicating that the predominant part of the water issued by steam holes and mud craters, as well as the steam supply of hot ground, is of meteoric origin. The case of the high-temperature fumaroles is more obscure. They may issue juvenile water in some quantity.

The chemical composition of natural steam does not vary greatly. Water is the main component, generally >95% by volume. With relatively few exceptions, CO₂ makes up >80% by volume of the gases intermixed with the steam. Typical examples of the chemical composition of the gases are given in Table 1.

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Cross-references: *Geyser; Hot ground; Hot spring; Solfatara; Thermal activity; Volcanoes and volcanism.*

G

GABBRO AND NORITE

Gabbro and *norite* are basic intrusive rocks composed principally of calcic plagioclase and pyroxene, with or without olivine. In gabbro the dominant ferromagnesian mineral is clinopyroxene, whereas it is orthopyroxene in norite. The expression “gabbroic rocks” includes both gabbro and norite. Gabbro is the coarse-grained, plutonic equivalent of basalt and dolerite.

Mineralogical composition

In most gabbros and norites, plagioclase predominates over pyroxene. In composition the plagioclase varies from labradorite to bytownite. Plagioclase with greater than 50% anorthite distinguishes these rocks from diorite, while the term *eucrite* is sometimes used for rocks which contain a very calcic plagioclase (bytownite to anorthite).

The Ca-rich pyroxene in these rocks is augite (usually diallage), in which the molecular proportion $Wo:En:Fs$ may vary from 45:45:10 to about 35:30:35. The orthopyroxene has a composition between En_{80} and En_{50} , i.e., Fe-rich bronzite and hypersthene. In many gabbroic rocks the hypersthene is not a product of primary crystallization, but has originated by inversion of pigeonite.

According to the relative proportions of cumulus plagioclase and pyroxene present, the gabbroic rocks can be subdivided as in Fig. 1. For rocks that contain both augite and hypersthene in nearly equal proportions, the term “hypersthene gabbro” has been used in the past. The IUGS Subcommission on the Systematics of Igneous Rocks recommends the name *gabbro-norite* for such rocks (Streckeisen, 1974), whilst Irvine (1982) suggests that they be referred to as either noritic gabbro or two-pyroxene gabbro, depending whether orthopyroxene or clinopyroxene, respectively, predominate.

Olivine and magnetite are common in gabbroic rocks, and if they are of a cumulus nature, the designations *olivine gabbro* and *magnetite gabbro*, respectively, are used. *Troctolite* is a gabbro composed of plagioclase and olivine with little or no pyroxene, whereas

allivalite is a troctolite in which the plagioclase is calcic bytownite and anorthite.

Rocks in which the ferromagnesian minerals are enriched in iron, such as ferroaugite, ferrohypersthene and Fe-rich olivine were previously known as “ferrogabbro.” However, since the accompanying plagioclase is usually andesine, the designation *ferrodiorite* for these rocks is more appropriate (Wager and Brown, 1968).

Other minerals which may be present in small proportions are quartz, apatite, biotite, and sulfides.

The term *hornblende gabbro* should only be used for gabbroic rocks in which the principal ferromagnesian constituent is primary hornblende, although it may be difficult to distinguish these rocks from those in which the amphibole is due to alteration of the pyroxene.

Gabbroic rocks can also be described in terms of decreasing abundance of cumulus minerals present (Irvine, 1982). A gabbro-norite could, for example, be a plagioclase–augite–hypersthene cumulate; olivine gabbro would be a plagioclase–olivine–augite cumulate; and so on.

Texture

Typical gabbroic rocks are hypidiomorphic granular. They are mostly cumulates, so that their texture is to a large extent determined by the postcumulus processes that were in operation. As a result, all types of cumulates, from adcumulates to orthocumulates are encountered (Wager et al., 1960).

Plagioclase is a cumulus phase in all the gabbroic rocks, usually as lath-like crystals which are often, especially in layered intrusions, orientated parallel to the plane of layering. Pyroxenes, olivine, and magnetite can either be cumulus or intercumulus. Quartz and biotite, when present, are always intercumulus.

Reaction between cumulus minerals (e.g., olivine and plagioclase) or between cumulus minerals and interstitial liquid is common. This gives rise to kelyphytic rims around the olivine grains, often associated with symplectite. Myrmekitic textures between cumulus plagioclase crystals are unique to the gabbroic rocks of the Bushveld Complex (Wager and Brown, 1968).

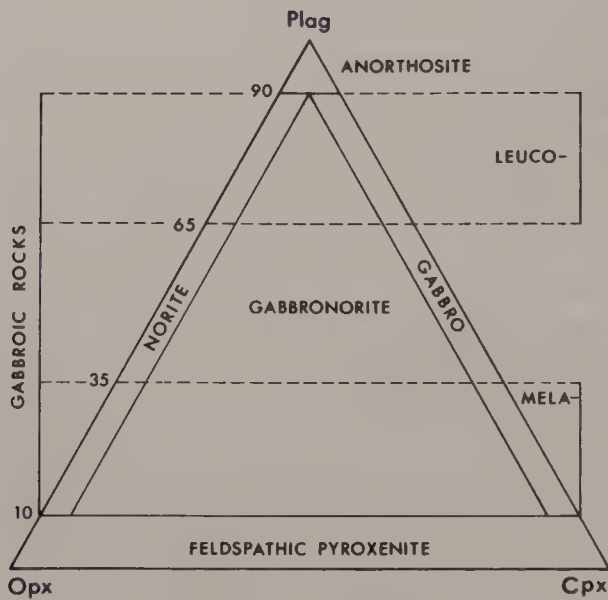


FIGURE 1. Schematic classification of gabbroic rocks: Plag, plagioclase; Opx, orthopyroxene; Cpx, clinopyroxene. (After Streckeisen, 1974.)

Chemical composition

Chemical analyses of gabbroic rocks are given in Table 1, together with the modes. The rocks contain $<54\%$ SiO_2 , which is an indication that they are basic. They have relatively high CaO and Al_2O_3 values, reflecting the high plagioclase content. The main chemical difference between gabbro and norite is the higher CaO content of gabbro, especially that which is in excess of the amount required for the plagioclase. The Niggli value $c-(al-alk)$ is accordingly a direct measure of the norite or gabbro affinity of a rock (Table 1). Seeing that the clinopyroxene in these rocks is augite, the amount of hypersthene in the norm is usually higher than that in the mode. Small amounts of either quartz or olivine are always present in the norm, whereas these minerals are often absent in the mode.

Occurrence and petrogenesis

Gabbro and norite are the normal products of the middle stages of fractionation of a basaltic magma. As a result, they are typical of layered intrusions but are also encountered in thick differentiated sills and smaller intrusions (Wager and Brown, 1968).

Irvine (1970) concluded, from theoretical considerations and experimental data on specimens from chilled margins of several layered intrusions, that plagioclase-clinopyroxene-olivine cumulates (olivine gabbros) should be the normal rock types of the middle stages of fractionation. The presence of plagioclase-orthopyroxene cumulates (norite and gabbro-

norite) is interpreted as being due to either a high load pressure during crystallization, or to the contamination of the magma by sialic material. In the latter case, assimilation of sialic material by basaltic magma will result in an increase in the amount of anorthite and magnesian pyroxene at the expense of Ca-rich pyroxene (Bowen, 1928).

The partial pressure of oxygen largely influences the trend of fractionation during crystallization of a basaltic magma and also determines the type and composition of the ferromagnesian phases of the reaction series (Osborn, 1962). Magnetite gabbro crystallizes under conditions of higher partial pressure of oxygen than gabbro which contains Fe-rich olivine and pyroxene.

Other types

Allanite—an altered gabbro containing actinolite and saussurite as euhedral pseudomorphs after the original minerals and maintaining the original ophitic texture (Allalin, Switzerland).

Bojite—a hornblende gabbro in which primary hornblende is the dominant mafic constituent; some augite or biotite may be present.

Corsilite (corsilyte)—an altered gabbro composed largely of smaragdite replacing diallage, and saussuritized feldspar.

Corsite—an orbicular gabbro.

Euphotide (eupholite)—euphotide was originally synonymous with gabbro, but the rock to which the term was first applied contained saussuritized feldspar; a talc-bearing euphotide is a eupholite.

Evjite—a hornblende gabbro like bojite in which the only light-colored mineral is labradorite or bytownite.

Gabbride—a field term for any igneous rock, such as gabbro and norite, with pyroxene as the only dark mineral and constituting $>50\%$ of the rock.

Ilmenite—a facies of ilmenite-bearing norite that consists predominantly of ilmenite.

Issite—a melanocratic hypabyssal rock composed chiefly of hornblende with smaller proportions of green pyroxene and labradorite; a melanocratic hornblende gabbro or a feldspathic hornblendite (Issa, U.S.S.R.).

Kedabekite—an igneous rock similar to eucrite composed of bytownite (originally identified as anorthite), Ca-Fe garnet and hedenbergite (Kedabek, U.S.S.R.).

Mareugite—a gabbro consisting mainly of bytownite and hauyne with variable proportions of hornblende and augite (Mareuge, France).

Muscovadite—a cordierite-biotite norite.

Ossipite (ossypite)—a troctolite composed of

TABLE 1. Chemical analyses and modal compositions of gabbroic rocks

	1	2	3 ^a	4	5	6
	Modes (vol. %)					
Plag	55.0	60.0	57.0	62.3	56.3	63.7
Ol	17.5	—	—	—	—	—
Opx	5.0	2.0	16.3	2.2	14.4	35.2
Cpx	21.0	33.0	26.7	35.5	29.3	1.1
Mt	1.5	5.0	—	—	—	—
	Chemical analyses					
SiO ₂	46.37	48.15	51.12	50.09	52.28	52.20
TiO ₂	0.79	2.64	0.19	0.14	0.15	0.08
Al ₂ O ₃	16.82	18.02	16.34	17.49	16.79	18.97
Fe ₂ O ₃	1.52	2.52	1.09	0.48	0.45	0.62
FeO	10.44	9.50	5.84	6.04	6.88	7.07
MnO	0.09	0.12	0.15	0.16	0.18	0.18
MgO	9.61	5.25	8.49	9.11	8.62	8.97
CaO	11.29	10.17	13.63	12.20	11.62	9.00
Na ₂ O	2.45	3.46	2.18	2.04	2.49	2.18
K ₂ O	0.20	0.14	0.06	0.18	0.18	0.21
H ₂ O +	0.38	0.22	0.56	1.57	0.73	0.84
P ₂ O ₅	0.06	0.05	0.03	0.01	0.01	—
CO ₂	—	0.03	—	0.33	0.09	0.07
TOTAL	100.02	100.27	99.68	99.84	100.47	100.39
c-(al - alk)	9.6	8.4	15.8	10.9	11.6	1.7

^a Mode given in wt. %.

1. Olivine gabbro (plagioclase-olivine-augite cumulate), Skaergaard intrusion; Wager and Brown (1968).

2. Gabbro (plagioclase-augite-magnetite cumulate), Skaergaard intrusion; Wager and Brown (1968).

3. Hypersthene gabbro (plagioclase-augite-pigeonite cumulate), Stillwater intrusion; Hess (1960).

4. Gabbro (plagioclase-augite cumulate), Bushveld Complex.

5. Hypersthene gabbro (plagioclase-augite-hypersthene cumulate), Bushveld Complex.

6. Norite (plagioclase-pigeonite cumulate), Bushveld Complex.

4,5,6 from unpublished data of D. R. Bowes and T. G. Molyneux.

labradorite and olivine with magnetite and a small proportion of diallage.

Ouenite—a eucrite-like rock containing green augite, anorthite and small proportions of hypersthene and olivine (Ouen, New Caledonia).

Rodingite—a medium-coarse-grained gabbroic rock containing grossular and diallage as essential minerals and generally Ca-enriched due to metasomatism; altered varieties contain prehnite and, or, serpentine.

Tilaite—a dark-colored plutonic rock intermediate in composition between gabbro and peridotite with abundant green diopside and olivine and a small proportion of calcic plagioclase (Tilai Kamen, U.S.S.R.).

Troutstone—a troctolite characterized by a speckled appearance due to dark spots of variably serpentinized olivine amongst light-colored feldspar.

Urbainite—an ilmenitite containing 10–20% rutile and 3–5% sapphirine (St. Urbain, Quebec, Canada).

GERHARD VON GRUENEWALDT

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Cross-references: *Basalt*; *Basic igneous rocks*; *Bushveld Complex*; *Cumulate textures*; *Dolerite*; *Intrusion-layered*; *Plutonic rocks—classification and nomenclature*.

GASES—VOLCANIC—see Vol. IVA

GEOCHEMICAL EVOLUTION OF THE CORE, MANTLE, AND CRUST—See Vol. IVA

GEOCHEMICAL MODELING IN PETROGENESIS

Geochemical modeling, as distinct from isotopic analysis, can be used to constrain source areas for magma generation or to distinguish groups of rock types, e.g., S- and I-type granites (Pitcher, 1979). Graphical or mathematical modeling techniques can be used to predict (within limits) the evolution of trace-element concentrations in magmas that have been modified by fractional crystallization, or they may be used to predict the composition of a melt derived by partial melting. Geochemical analyses are normally present in the form of weight percent (wt.%) oxides for major elements (for cations like Fe^{2+} , Mg^{2+} , K^+), in parts per million (p.p.m.) for trace elements (e.g., Rb, Sr, Ga) and in p.p.m. on chondrite-normalized plots for rare-earth elements (REE). This information, treated either separately, or together, means that all three groups can contribute to the development of a petrogenetic model that may be used to explain geochemical trends observed in rock suites.

The simplest form of geochemical modeling is the representation of a changing magma composition on a variation diagram (Fig. 1a,b; cf., Wilcox, 1979). This form of presentation allows a changing magma composition to be interpreted with reference to the addition or subtraction of material of known composition. The concentration of a particular oxide in a magma, modified either by subtraction or addition of that oxide, will vary along the line joining the initial concentration of the oxide in the magma with the concentration in the added or subtracted material by an amount propor-

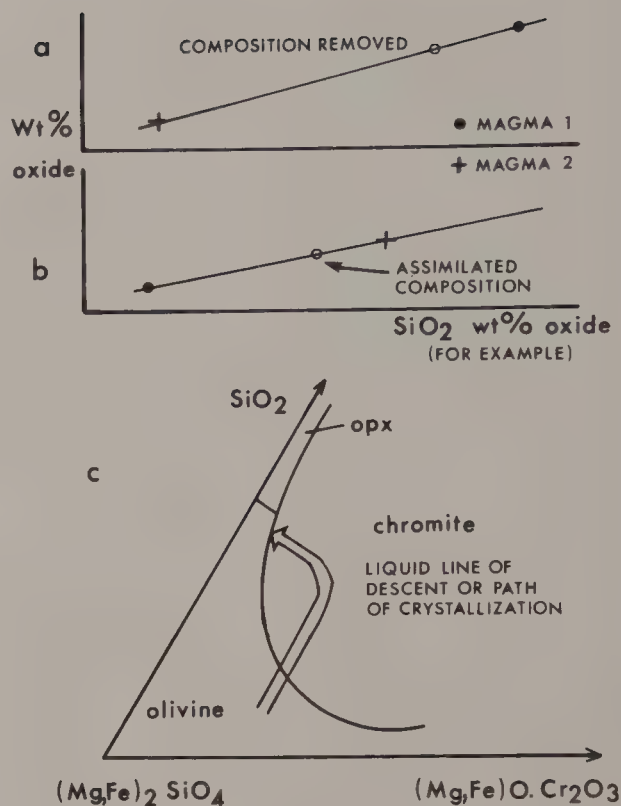


FIGURE 1. (a),(b) Hypothetical variation diagrams showing the relationship between two magmas and (a) a fractionally crystallized composition and (b) an assimilated composition. (c) Schematic projection from $(\text{Mg}, \text{Fe})\text{O}$ onto the plane SiO_2 – $(\text{Mg}, \text{Fe})_2\text{SiO}_4$ – $(\text{Mg}, \text{Fe})\text{O} \cdot \text{Cr}_2\text{O}_3$, showing a hypothetical crystallization path entering the chromite field after the addition of SiO_2 . (After Irvine, 1977.)

tional to the quantity added or subtracted. The geological processes of fractional crystallization or partial melting (cf., White and Chappell, 1977) may be analogous to subtraction, whereas assimilation would be analogous to addition.

A striking example of the effects of assimilation is provided in the Muskox intrusion, Canada. The assimilation of a silica-rich granophyre present in the country rock, not only had the effect of increasing the proportion of SiO_2 in the ultrabasic portion of the magma, but was ultimately responsible for the precipitation of the chromite that formed chromitite layers. The displacement of the magma composition into the primary crystallization space of chromite in the projection SiO_2 – $(\text{Mg}, \text{Fe})_2\text{SiO}_4$ – $(\text{Mg}, \text{Fe})\text{O} \cdot \text{Cr}_2\text{O}_3$ (Fig. 1c; cf., Irvine, 1977) by the addition of silica shows how cryptic variations in chemistry are often expressed petrographically.

The removal of certain elements from a melt by the fractional crystallization of a phase may adequately explain subtraction of certain elements over a particular range of concentrations, but the process cannot proceed indefin-

itely, since the continual modification of a melt's composition would lead to the exhaustion of the elements being removed. In addition, a particular element may be removed from a magma simultaneously by two or more crystallizing phases. Only rarely would one phase crystallize totally independently of any other, except perhaps in the development of olivine or pyroxene cumulates and anorthosites.

The evolving trace-element concentration in a magma can be modeled for fractional crystallization using a Rayleigh equation,

$$\frac{C_1}{C_0} = (1 - F)^{D-1}$$

where C_1 is the concentration of an element in the melt, C_0 is the starting concentration of an element in the parent magma, F is the fraction of liquid remaining, and D (the bulk distribution coefficient) is given by

$$D = \sum_{i=1}^n Kd_i w_i,$$

where Kd_i is the distribution coefficient of element in phase i , w_i is the weight fraction of phase i , and

$$Kd = \frac{\text{concentration of element in mineral}}{\text{concentration of element in melt}}.$$

The Rayleigh equation is of necessity an approximation of real open-system fractionation, since it assumes the precipitation of an infinitesimally small amount of a phase accompanied by its immediate removal from any further interaction with the liquid. This is at variance with known and observed petrographical features resulting from the resorption of crystallized phases and the intense, localized geochemical gradients expected during supercooling and the development of spinifex textures. This modeling technique, being limited to a closed system, cannot realistically accommodate simultaneous fractional crystallization and assimilation of surrounding material, or the introduction of more "fresh" magma. A refinement of this problem was proposed by O'Hara (1977), who modeled the geochemical variations produced by a periodically refilled magma chamber.

A major constraint on which phases will fractionally crystallize from a melt must be the observed petrography of the rocks that form as end products of the process. On textural grounds it may be possible to suggest that the development of cumulate or porphyritic textures approximates closest to Rayleigh fractionation. Assuming that this process has been operative, a geochemical trend may be explained in terms

of a fractional crystallization vector or vectors (calculated on the basis of several crystallizing phases in known proportions) using a bulk distribution coefficient. An alternative presentation would be to assume that only one phase crystallized, in order to assess its influence on the development of an observed trend (in this case it is only necessary to consider the distribution coefficients for one phase assuming it constitutes 100% of the crystallizing phases). This may show, for example in the case of Fig. 2, that plagioclase and K-feldspar had the greatest influence on the evolution of the magma from which the granites crystallized, and that the fractional crystallization of small amounts of biotite and/or hornblende might be used to explain any lateral variations from the trend. Noticeably, biotite appears to have had a significantly greater influence on the granodiorite trend.

This method can be used (1) to test whether the concentration of a given element in the parent magma was a realistic starting point and (2) whether (for two elements plotted) another

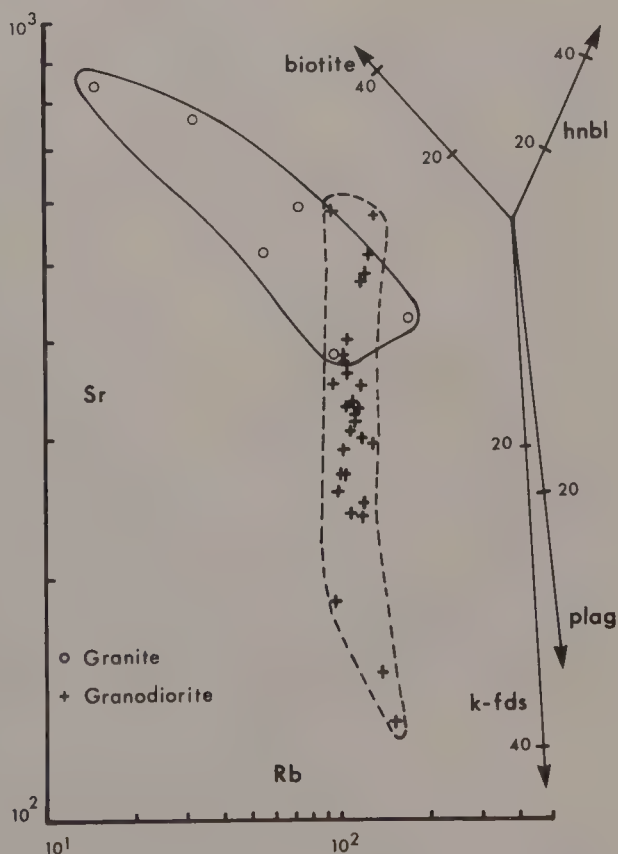


FIGURE 2. Log Sr-log Rb plot for granites and granodiorites from E Finland, in relation to fractional crystallization vectors determined for biotite, plagioclase (plag), K-feldspar (k-fds), and hornblende (hnbl). The percentage of fractional crystallization of an original composition to produce comparable changes in Rb and Sr contents is marked along the vectors.

phase, perhaps not observed in the rock under consideration, had a more significant effect on the trend. It may also be possible, using several element plots, to suggest a hypothetical parent magma composition from which fractional crystallization could proceed on the basis of inferred starting points for the individual mineral vectors.

A fractional crystallization process may explain the evolution of a liquid once it has been derived, but the generation of liquids of different composition may initially be achieved by partial melting, either of different sources or by various degrees of partial melting of a single source. These processes may again be modeled on a variation diagram, but more quantitative estimates of the concentration of an element in a liquid, aimed at isolating possible source material, may be determined by the application of the formula

$$\frac{C_1}{C_0} = \frac{1}{F + D - FD}$$

The formula represents "batch melting"; the melt is removed from its coexisting residue in a single batch so as to prevent any further interaction. This approach, however, limits the resolution of individual phases of anatexis, and again is likely to be an oversimplification of any natural process. In order to make any model more rigorous, knowledge of the mineralogy of any potential source rock would be required in conjunction with an estimate of the heat supplied to the rock, in order to assess which phases may have melted (cf., Allègre and Minster, 1978). Knowledge of the phase relations in the assemblage under consideration is also required to determine whether it will melt congruently or incongruently. Both these considerations are necessary because during either progressive anatexis or in different phases of anatexis, D (the bulk distribution coefficient) would be expected to change.

A more accurate estimate of changing liquid compositions, considered in the light of different bulk distribution coefficients and several phases of anatexis, could be determined by the formula

$$\frac{C_{1,2}}{C_0} = \frac{D_1}{(F_1 + D_1 - F_1 D_1)(F_2 + D_2 - F_2 D_2)},$$

representing incremental batch melting (the subscript 2 indicates conditions during the derivation of the second liquid, and the concentration of an element in a coexisting residue may be determined from $C_{\text{RESIDUE}} = D \cdot C_{\text{LIQUID}}$).

REE data may be manipulated in a fashion similar to that described for trace elements, using partition coefficients. Normally, however,

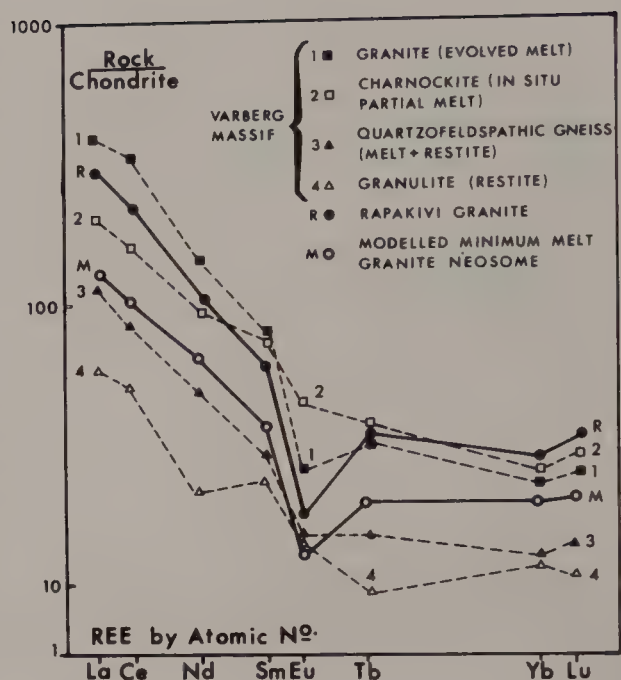


FIGURE 3. Distribution of REE modeled for the rocks of the Varberg massif, Sweden. The granite (1), whose pattern shows a striking resemblance to a Finnish rapakivi granite (apart from differences in the Eu anomaly), is interpreted as resulting from the migration of the in situ partial melt and its subsequent evolution. (After Hubbard and Whitley, 1978.)

the REE data for a particular rock suite are presented in the form of a chondrite-normalized plot or concentration vs. atomic number. Interpretations based on such information are generally of such form that the geochemical evolution of a suite of rocks represents either light (or heavy) REE enrichment or depletion, assessed on the overall gradient of the line joining the analysis data points (Fig. 3).

Given that light REE generally behave in a fashion similar to incompatible elements with values of $K_d < 1$, then partial melting would normally be expected to result in light REE depletion in the source and enrichment in the liquid; this emphasizes the necessity of field control on the samples analyzed, as that is where the initial judgment may have been made whether a rock represents a source or a derived liquid. Coexisting with light REE enrichment in the liquid would probably be heavy REE enrichment in the restite (noting that these are relative terms in the light of an unknown number of phases of anatexis). For example, garnet, considered to be a major component of mantle restite after about 20% partial melting, preferentially retains heavy REE. This is an illustration of the influence of crystallo-chemical controls on the absolute values of distribution coefficients.

A common feature observed on many REE plots is a Eu anomaly that indicates the incorporation of Eu in plagioclase during fractional crystallization in preference to other REE. The anomaly is expressed as a peak or a trough on the plot, depending upon whether plagioclase has been preferentially accumulated or removed; this again illustrates that geochemical anomalies, often cryptic, may have a conspicuous mineralogical expression (Fig. 3).

Modeling a geochemical trend using the interrelationship between two elements (i.e., in two dimensions) is a limiting perspective on the evolution of a suite of rocks. Ideally, multi-dimensional space, determined by a number of variables operative during open system fractionation, would be required to produce an accurate model. A more reasonable approach to this problem is the reduction of the number of variables to those major elements required by the main mineral constituents of the rock suite being analyzed—e.g., CaO , MgO , SiO_2 , and Al_2O_3 , which constitute the CMAS system that is commonly used to model the relationship between mantle compositions and quartz-saturated basalts. Such reduction assumes that FeO , K_2O , and Na_2O are not major components in the phases involved.

Data are normally presented in the form of triangular plots projected from a phase present in the rock suite being analyzed (Fig. 4) in order to determine how the liquid composition evolves with respect to the other three variable components. If, for example, in the case of a projection from clinopyroxene, a liquid fractionally crystallized orthopyroxene, then the evolving liquid composition would trend away from the orthopyroxene coordinates, i.e., MgO and SiO_2 would simultaneously be removed from the liquid in equal proportions. This produces a "liquid line of descent" that plots the evolving concentration of the components defined in the projection during their preferential incorporation in a crystallizing phase; alternatively, the plot produced can also be referred to as the path of crystallization. If, during the development of a melt, physical conditions changed within a crystallizing body of magma in terms of pressure, temperature and/or composition that initiated the crystallization of another phase (e.g., clinopyroxene), then a projection from SiO_2 may be required to elucidate this.

Complex mathematical models have led to the development of computer models of geologic processes that endeavor to calculate

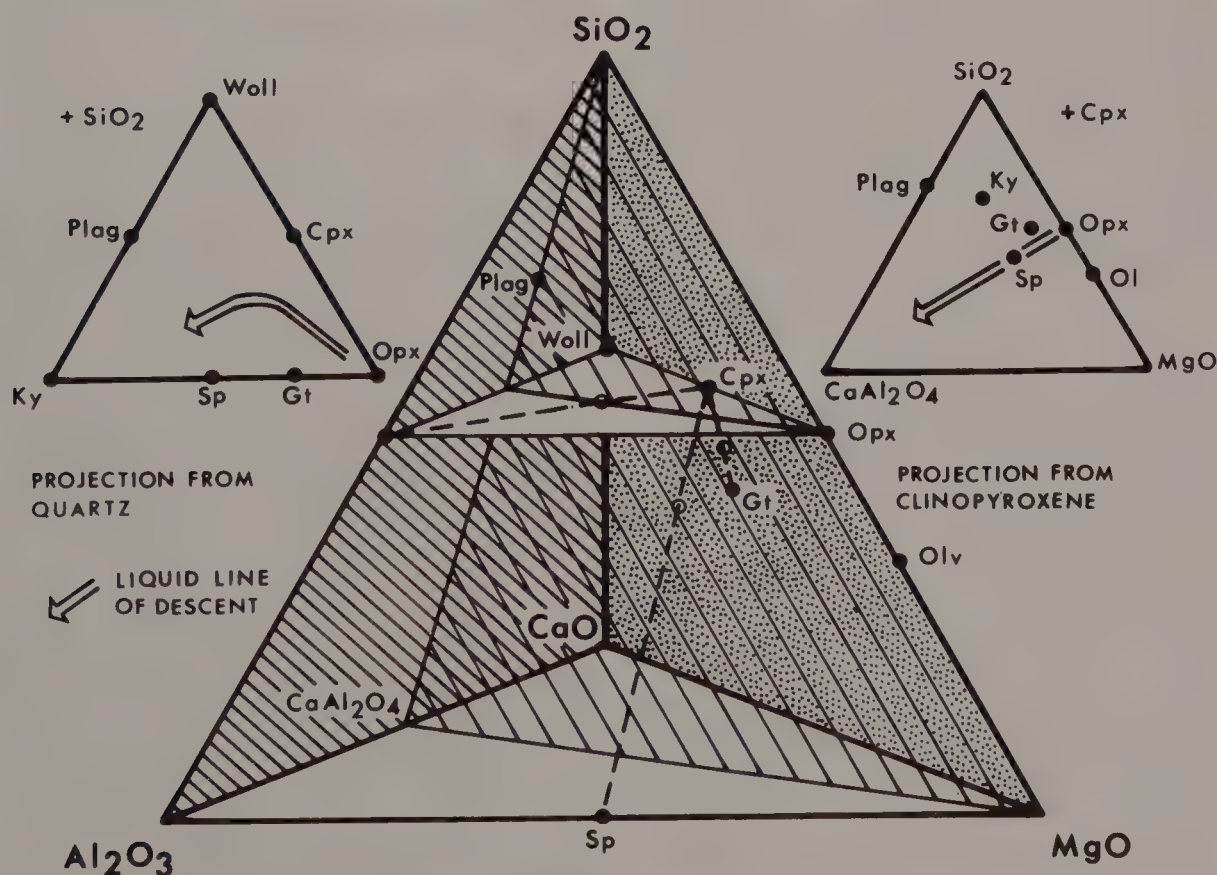


FIGURE 4. Schematic tetrahedron representing the CMAS system indicating the position of various minerals within the space (●); dotted lines represent projections from clinopyroxene, and open circles (○) represent their intersection with the plane SiO_2 - CaAl_2O_4 - MgO shown as an equilateral triangle to the right of the tetrahedron.

the proportions of added or subtracted material (of known composition, ideally) that would lead to the closest correspondence of two observed analyses assuming that they represent magmas in an evolving sequence. An example of one such modeling program is XLFRAC (Stormer and Nicholls, 1977). This program compares the bulk composition (in terms of major elements) of two analyses assumed to have evolved by the fractional crystallization or assimilation of one or more phases (if possible, analyzed from the proposed "final magma"). It calculates the probable amounts of the phases that would have to be added or subtracted to or from an "initial melt" to give an observed difference between it and the "final magma." Good "fits" are judged when there are only very low concentrations of residual components left and these can only be achieved for all components with a thorough knowledge of the petrography of the final magma (rock).

By definition, modeling techniques can only approximate to natural systems, as they attempt to simplify and reduce the number of variables to be considered. They concentrate on the major components in a rock that appear to have had the greatest influence on development, and the major limitations may be centered on a reasonable form of presentation and a lack of physical data concerning the effect of one variable or another. Should a graphical solution correspond to an observed trend, the limited dimensions must be taken into consideration. Furthermore the thermodynamic suitability of a solution must be tested in relation to the observed petrography of the rock suite considered, bearing in mind that the original sample may be imperfect.

N. M. HALDEN

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Cross-references: *Assimilation; Differentiation index; Double-diffusive convection; Element partitioning in petrogenesis; Eutectic melting; Hybridization; Magma; Magmatic crystallization in open systems; Magmatic differentiation; Migmatite—origin of neosome; Petrochemical calculations.*

GEOCHEMISTRY—See Vol. IVA

GEOCHRONOMETRY—See Vol. IVA

GEOSYNCLINAL VOLCANISM

The character of geosynclinal volcanic rocks is of great interest as a possible indication of the nature of crustal structure and thickness beneath geosynclines, and of their tectonic setting. According to the classic views elaborated in the first half of this century (summarized by Tyrrell, 1955), geosynclinal volcanism in eugeosynclines (Kay, 1951) was held to be a distinctive and unique opening phase of a tripartite magmatic cycle in orogenic regions. The geosynclinal phase supposedly includes the eruption of mafic lavas, chiefly basalt and spilite, together with the emplacement of gabbro and peridotite or serpentinite into the lower levels of the growing geosynclinal accumulation. These mafic and ultramafic igneous rocks were collectively called ophiolites, now thought to represent ancient oceanic crust. Their composition and texture was held to contrast strongly with the granitic intrusive rocks of the later orogenic phase, and also with andesitic lavas and pyroclastic rocks that might be erupted by leakage from granitic magma chambers during orogenesis and later. These interpretations imply a regular geotectonic cycle

by which all geosynclinal belts pass from conditions of passive subsidence to tectonic deformation and uplift. They need modification because of the now abundant evidence that there is no standard geotectonic cycle but rather a number of variants representing different plate-tectonic regimes.

W. R. DICKINSON

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Cross-references: *Basalt*; *Granite*; *Igneous rocks—tectonic setting*; *Ophillite*; *Steinmann Trinity*; *Spillite*; *Ultramafic belts*.

GEYSER

A geyser is an intermittent hot spring throwing out columns of water and steam at more or less regular intervals; the term is derived from the Icelandic name for such springs (geysir). Geysers are found only in some thermal areas, mainly in Iceland, New Zealand, and in the Yellowstone National Park, Wyoming, U.S.A., where the greatest number is located.

The classical example is the Great Geysir in Haukadalur in Iceland, which appears to have been active throughout most of historic time (Fig. 1). This spring is located upon a hill 30–40 m in diameter and 1.5 m high made up of siliceous sinter (porous opaline variety of silica deposited from the hot waters). The center of the hill is a bowl-shaped depression, ca. 20 m in diameter, with a 3 m wide central circular hole above a pipe 20 m deep with a diameter of ca. 1 m at the bottom. During quiet periods between eruptions, 2.5 liters/second of water flow out of the pipe and over the rim of the depression. The temperature of the outflowing water is 70–90°C depending on weather conditions; at the bottom of the pipe it is ca. 125°C. Convection in the upper part of the pipe causes temperature fluctuations and cooling. An eruption may be initiated in calm weather when there is less surface cooling. Violent convection in the upper part of the pipe, and underground shocks resulting from collapsing steam bubbles, are the first signs of the beginning of an eruption. The surface of the water rises in a turbulent fashion, the height increasing slowly until the eruption is at maximum intensity. During the main phase, a column of water and steam is thrown to heights

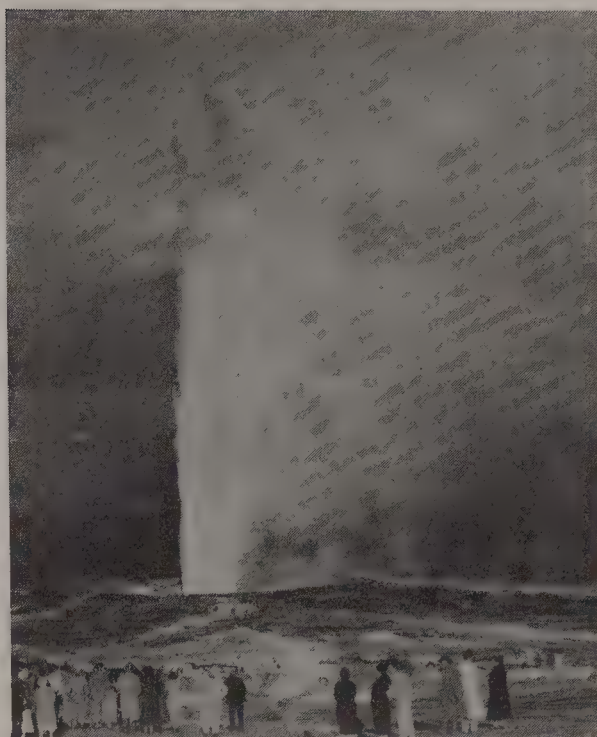


FIGURE 1. The Great Geysir in Haukadalur, Iceland.

of 40–60 m. The eruption may last for 10–15 minutes, during which the water content of the column is gradually depleted ending in an eruption of only slightly wet steam. The pipe appears to have emptied itself of water at the end of the eruption. Some time elapses before it is gradually filled again. The activity of the Great Geysir is irregular. During favorable conditions in the past, eruptions occurred perhaps every day, but the activity has decreased and eruptions have now to be initiated by soap, which is thrown into the pipe.

There are ca. 30 smaller geysers in Iceland. In Yellowstone National Park there are about 200, with one in ten of the springs being a geyser. Old Faithful is the best-known; it erupts at regular intervals of about 1 hour, and sends a column of steam and water 30–50 m into the air for 5 minutes. Other well-known geysers in the Yellowstone Park are the Castle Geyser, Grand Geyser, and Beehive Geyser. The latter erupts to about 70 m, but is irregular.

The most powerful known geyser was the Waimangu Geyser on the North Island of New Zealand. It was active during the period 1900 to 1904 and had a maximum height of eruption of about 500 m.

Geyser action is a special type of relaxation oscillation. In most cases, water having a temperature of 120–150°C enters the lower part of a vertical pipe where the hydrostatic pressure is sufficient to prevent boiling. The water

ascends in the pipe and is subjected to gradual cooling, mainly because of convection. When this cooling is sufficient to prevent boiling anywhere in the pipe, the water may flow quietly out at the surface. Some geysers do not flow during periods between eruptions. However, the temperature conditions are unstable owing to the fluctuating character of the convection and there are intervals during which the convective cooling is insufficient to prevent boiling and the formation of steam bubbles. Gases in solution in the water may also contribute to the formation of bubbles. Some steam bubbles collapse again, but other bubbles may be large enough to throw out a part of the water in the upper pipe.

This results in a lower hydrostatic pressure in the entire pipe and initiates violent boiling throughout the water column. A steam-water mixture is blown out with a great velocity. In many cases the velocity of the emerging steam phase is of the order of 100 m/second; the water phase has a considerably lower velocity owing to its greater inertia. Judged from the height of eruption the water droplets may in many cases be thrown out at a velocity of 30–50 m/second. As the eruption proceeds, the geyser pipe is gradually emptied, resulting in a lower hydrostatic pressure in the cracks and fissures feeding the pipe.

During the intervals between eruptions, the walls of the pipe and of the feeding fissures reach thermal equilibrium with the inflowing water and have approximately the same temperature as the water. Therefore, as the hydrostatic pressure is gradually lowered, the wall temperature becomes greater than the boiling temperature of the water remaining in the pipe. The great mass of rock in the walls therefore acts as a heat source contributing to further flashing of the remaining water. This may account for the final steam phase—the geyser eruption. The flashing lowers the temperature of the walls, resulting in a wall temperature of about 100°C as the eruption comes to an end and the inflowing water starts filling the pipe. This water, which has a temperature well above 100°C, is cooled upon contact with the walls and no further flashing occurs. The pipe is refilled quietly, and water may again flow out at the surface (the outlet is referred to as the bore). The heat supplied by the incoming water gradually raises the temperature of the wall rock, resulting again in an equilibrium after a certain interval of time. Conditions for a new eruption are created, and the whole process repeats itself.

The time interval required for the temperature equilibrium to be attained is the minimum period of eruption. This accounts for the

regularity of some geysers. The length of this period depends on the ratio between the heat supply and the thermal inertia of the rock mass that has to be brought into equilibrium with the incoming water. On the other hand, whether an eruption is initiated at the end of the minimum period depends on the convective cooling in the upper part of the pipe. Geysers with little convective cooling will erupt regularly at intervals equal to the minimum period. Geysers subjected to great cooling may erupt only irregularly at much longer intervals. Studies on the Great Geysir in Iceland have indicated that explosive boiling due to a superheated state of a portion of the upflowing water may be necessary in order to initiate an eruption. It appears as if the soap used to initiate eruptions may enhance the degree of superheat.

All geyser areas are characterized by a relatively high content of silica in the water, and deposits of siliceous sinter (geyserite) contribute to the building up of the geyser system.

G. BODVARSSON

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Cross-references: *Fumarole*; *Hot ground*; *Hot spring*; *Solfatara*; *Thermal activity*; *Volcanoes and volcanism*.

GLASS, DEVITRIFICATION OF VOLCANIC—

See Vol. IVB

GLAUCOPHANE SCHIST FACIES

A very distinctive facies characterized by blue schist containing abundant glaucophane is developed mainly in the circum-Pacific regions and the Alpine-Himalayan orogenic belt. The facies is distinguished by minerals and mineral assemblages stable under high-pressure and low-temperature conditions, viz., lawsonite, aragonite, and jadeite + quartz. The term glaucophane schist facies was proposed by Eskola (1939), but was redefined as

glaucophane-lawsonite schist facies by Fyfe and Turner (1966). Glaucophane schist facies has priority and is used here; it is synonymous with *blue schist* and *prasinite*, a term still in common usage in some European and Russian literature.

Rocks recrystallized in the glaucophane schist facies occur in the same tectonic zones as tectonic mélange, and the disrupted nature of such zones has often obscured the relationship of this facies to others. Such zones frequently contain blocks of eclogite, and ophiolitic debris, metabasalts, and spilites, pelagic shale, and graywacke. Where glaucophane schist facies assemblages grade into greenschist facies several relationships have been described, most commonly one of greenschist overprinting, though a transitional *lawsonite-albite-chlorite facies* has been described. The relationship to eclogite is also equivocal, with both prograde and retrograde relationships recorded.

Internal subdivision of this facies into subfacies and zones has also proved equivocal, although Coleman and Lee (1963) devised a scheme based on Californian metabasalts, and Chopin and Schreyer (1983) have done the same for metapelitic rocks.

Mineral assemblages

Metapelites. Chlorite, quartz, pyrophyllite, talc, kyanite, and garnet are the typical phases contained in metapelites in this facies. Few such assemblages are diagnostic, though the replacement of Mg-chlorite by kyanite and talc does indicate high pressures. The presence of carpholite $((\text{Mg}, \text{Fe}^{2+})\text{Al}_2\text{Si}_2\text{O}_6(\text{OH})_4)$ and Mg-rich chloritoid $((\text{Mg}, \text{Fe}^{2+})\text{Al}_2\text{OSiO}_4(\text{OH})_2)$ is considered diagnostic (Chopin and Schreyer, 1983). The distribution of these two new phases permits further subdivision (Fig. 1).

Metacarbonates. Aragonite, the high-pressure polymorph of CaCO_3 , is the distinctive mineral of carbonates recrystallized in this facies. At low temperatures it is only stable above 5–10 kb and is metastable at the surface. In exposed blue schists it has often been completely or partially replaced by inversion to calcite. Other distinctive, but not diagnostic, minerals include pyralspite garnet, calcic amphiboles, stilpnomelane and clinozoisite.

Metagraywackes. The association of graywackes with glaucophane schist terrane is ubiquitous. Typical minerals developed include quartz, jadeite, lawsonite, chlorite, stilpnomelane, muscovite, aragonite, and glaucophane. In many cases, origin detrital textures are preserved up to the highest grades of metamorphism in this facies.

Metabasalts. Glaucophane is widespread and distinctive in metabasalts, particularly those

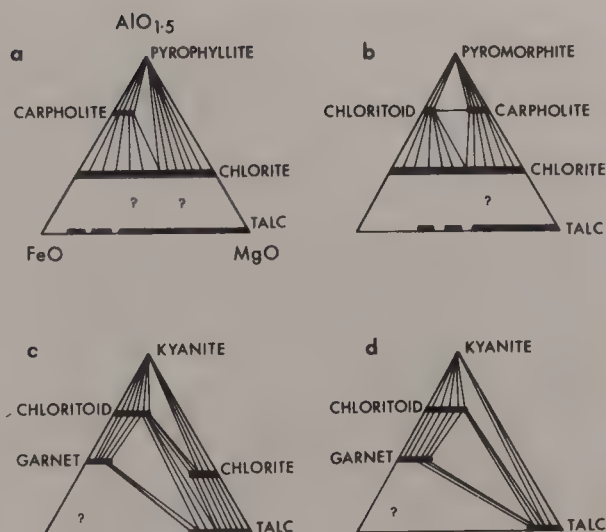


FIGURE 1. Phase relations in metapelites in the glaucophane schist facies; (a)–(d) show increasing metamorphic grade, particularly with reference to carpholite and magnesiochloritoid and represent four subfacies. (After Chopin and Schreyer, 1983.)

that are spilitized. In addition, these rocks also contain lawsonite, pumpellyite, epidote, jadeite, muscovite, chlorite, sphene, rutile, calcite–aragonite, and garnet. In California four distinct subfacies have been defined on the basis of mineral assemblages in such rocks, ranging from virtually unaltered to gneissose in appearance (Table 1).

Metacherts. These are minor components of many terranes, but include a wide variety of exotic minerals and assemblages: deerite, howieite, and zussmanite appear to be diagnostic of this facies (Lathard and Schreyer, 1981). These three are, however, rare accessory phases. The main assemblages commonly contain essential quartz plus stilpnomelane, spessartine-rich garnet, epidote, aegerine, and piemontite (Mn-epidote).

Eclogite. A distinctive variety of eclogite occurs in glaucophane schist terranes, often as blocks in tectonic mélange. It is an omphacite–almandine-rich garnet rock with abundant rutile. This assemblage is often replaced by more typical glaucophane schist facies assemblages like pumpellyite, lawsonite, glaucophane, and chlorite. Both retrograde and prograde relationships are recorded, though the overall relationship remains enigmatic.

P–T conditions

Experimental studies have shown that many of the distinctive minerals and mineral assemblages of this facies are only stable under high pressures and relatively low temperatures, i.e., with a heat flow less than that producing the average geothermal gradient. Glaucophane

TABLE 1. Mineral parageneses in metabasalts of types II, III, and IV in the blueschist (glaucophane schist) Franciscan Formation, California, U.S.A., (type I is unaltered basalt)

	Type II	Type III	Type IV
Glaucophane	-----	-----	-----
Lawsonite	-----	-----	-----
Pumpellyite	-----	-----	-----
Clinozoisite	-----	-----	-----
Epidote	-----	-----	-----
Jadeitic pyroxene	70-80% Jd	70-80% Jd	30-40% Jd
Muscovite	1M	2M ₁	2M ₁
Chlorite	-----	-----	-----
Aragonite	-----	-----	-----
Calcite	-----	-----	-----
Sphene	-----	-----	-----
Rutile	-----	-----	-----
Garnet	?	Al-Gr-Sp	Al-Gr
Apatite	?	-----	-----
Pyrite	-----	-----	-----
Texture	Non-foliated	Foliated	Gneissose
Average density	2.88	3.05	3.20
H ₂ O wt.% average	4.0	3.1	1.5

Source: after Coleman and Lee (1963).
Jd, jadeite molecule; Al, almandine molecule; Gr, grossular molecule; Sp, spessartine molecule; ? indicates a mineral not observed but which may be present in small amounts.

may only be stable above 10 kb (Fig. 2; Carman and Gilbert, 1983), whereas lawsonite is only stable below ca. 400°C. Typical *P-T* estimates from glaucophane schist terranes generally fall into the range 8-18 kb and 200-500°C, suggesting thermal gradients below 10°C/km.

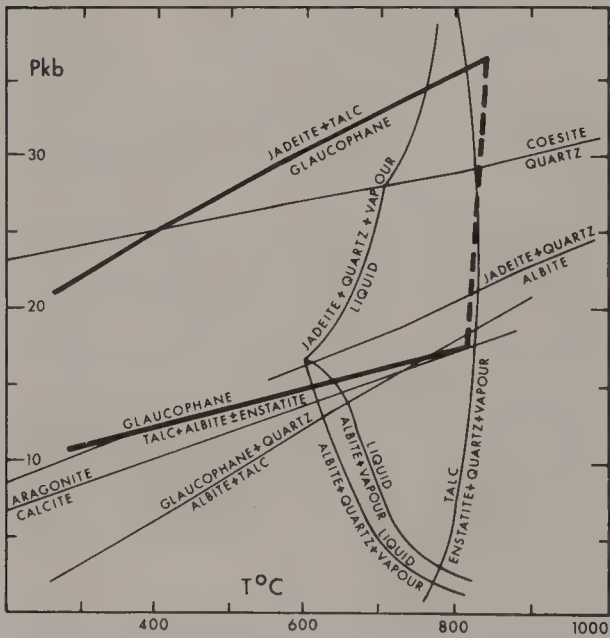


FIGURE 2. Petrogenetic grid for the glaucophane schist facies showing the stability field for glaucophane in relation to other important phases (aragonite, jadeite, albite, talc). (After Carman and Gilbert, 1983.)

Veins of metamorphic silicates are extremely common in some glaucophane schist terranes, suggesting hydrostatic fracturing, consistent with fluid pressure exceeding lithostatic pressure. The vein minerals are always Na-, Ca- and H₂O-bearing irrespective of host-rock composition. Transport in a fluid phase, and metasomatism, particularly involving Na, appear to be important mechanisms during metamorphism in this facies.

Occurrence and origin

The constraints of high pressure and low temperature (low heat flow) are best satisfied in those zones where subduction of oceanic crust (relatively cold) or tectonic thickening of continental crust are operative processes. Indeed, for some time, the occurrence of glaucophane has been regarded as unique evidence for subduction processes in the geological record. Glaucophane schist facies terranes are at present, and in the recent geological record, associated with active subduction zones or sites of continental collision—the circum-Pacific margin and the Alpine-Himalayan belt. Through time, most are Mesozoic or Cenozoic in age, with isolated Palaeozoic examples, and a few contentious examples from the Precambrian. This may be a function of uplift and erosion, or the overprinting of this facies by greenschist facies assemblages as heat flow recovered after subduc-

tion ceased. This temporal anomaly has been used to argue that plate tectonic processes were not operative in the Precambrian.

ADRIAN F. PARK

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Cross-references: *Eclogite*; *Eclogite facies*; *Greenschist facies*; *Island arcs—petrology*; *Lawsonite-albite-chlorite facies*; *Metamorphic facies*; *Paired metamorphic belts*.

GNEISS

Gneiss has been defined as a foliated rock formed by regional metamorphism in which (granulose) bands or lenticles of granular minerals alternate with (schistose) bands and lenticles of flaky or elongate prismatic minerals. Generally, <50% of the minerals show preferred parallel orientation. Although many gneisses are quartzofeldspathic, the mineral composition is not an essential factor in its definition (American usage). Varieties are distinguished by texture (e.g., augen gneiss), characteristic minerals (e.g., hornblende gneiss), structure (e.g., pencil gneiss composed of pencil-shaped quartz-feldspar aggregates mantled by mica flakes), or general composition and/or origin (e.g., granitic gneiss).

The most common use of the term “gneiss” is for a medium-coarse grained, generally equi-

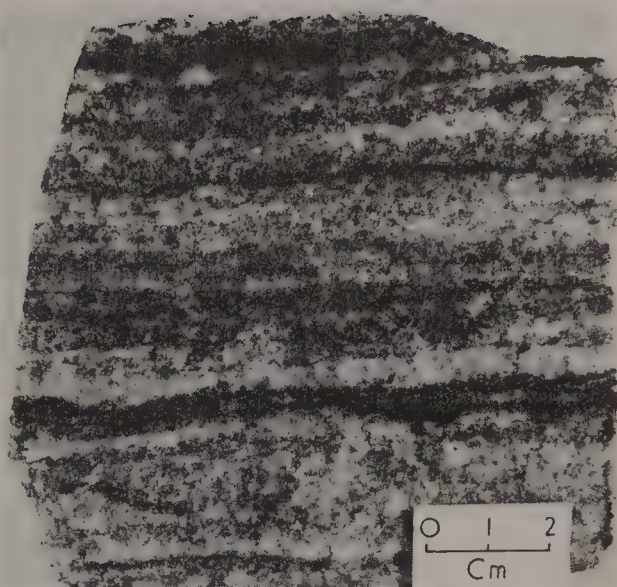


FIGURE 1. Garnet-sillimanite-biotite gneiss; dark layers rich in biotite and sillimanite alternating with lighter quartzofeldspathic bands; Kandy, Sri Lanka.

granular, metamorphic rock, with a poorly-defined planar anisotropy or gneissosity indicated by alternating felsic and mafic layers (Fig. 1). The thickness of these layers is variable; the mafic or dark layers are usually >0.5 mm, while the felsic bands tend to be thicker and can attain a thickness of 40 cm or more in banded gneisses (Katz, 1970). With diminishing thickness of the layers to fine laminae dominated by oriented minerals, the rock becomes a schist (Wenk, 1963).

Classification

Gneiss is an old German mining term first applied to rocks by miners of the Erzgebirge (Ore Mtns.—E Germany—Czechoslovakia), who used it to designate the country rocks near ore veins. It came into the English language with the translation of Henckel's *Pyritol* in 1757. Werner defined it in 1787. Since then it has become an established metamorphic rock name with several mineralogic, petrographic, structural, textural, and petrogenic varieties recognized. Gneisses are widely distributed in metamorphic belts of all ages and are most frequently encountered in Precambrian terranes. The most common varieties are defined on the basis of diagnostic metamorphic minerals in order of relative abundance (e.g., garnet-sillimanite gneiss is a gneiss containing garnet and sillimanite with the former predominating). Petrographic types often mentioned in the literature include granitic gneiss, tonalitic gneiss, gabbro-gneiss (e.g., zobtenite), etc. If structures are apparent, terms such as banded gneiss, smallfolded gneiss, and wavy gneiss have

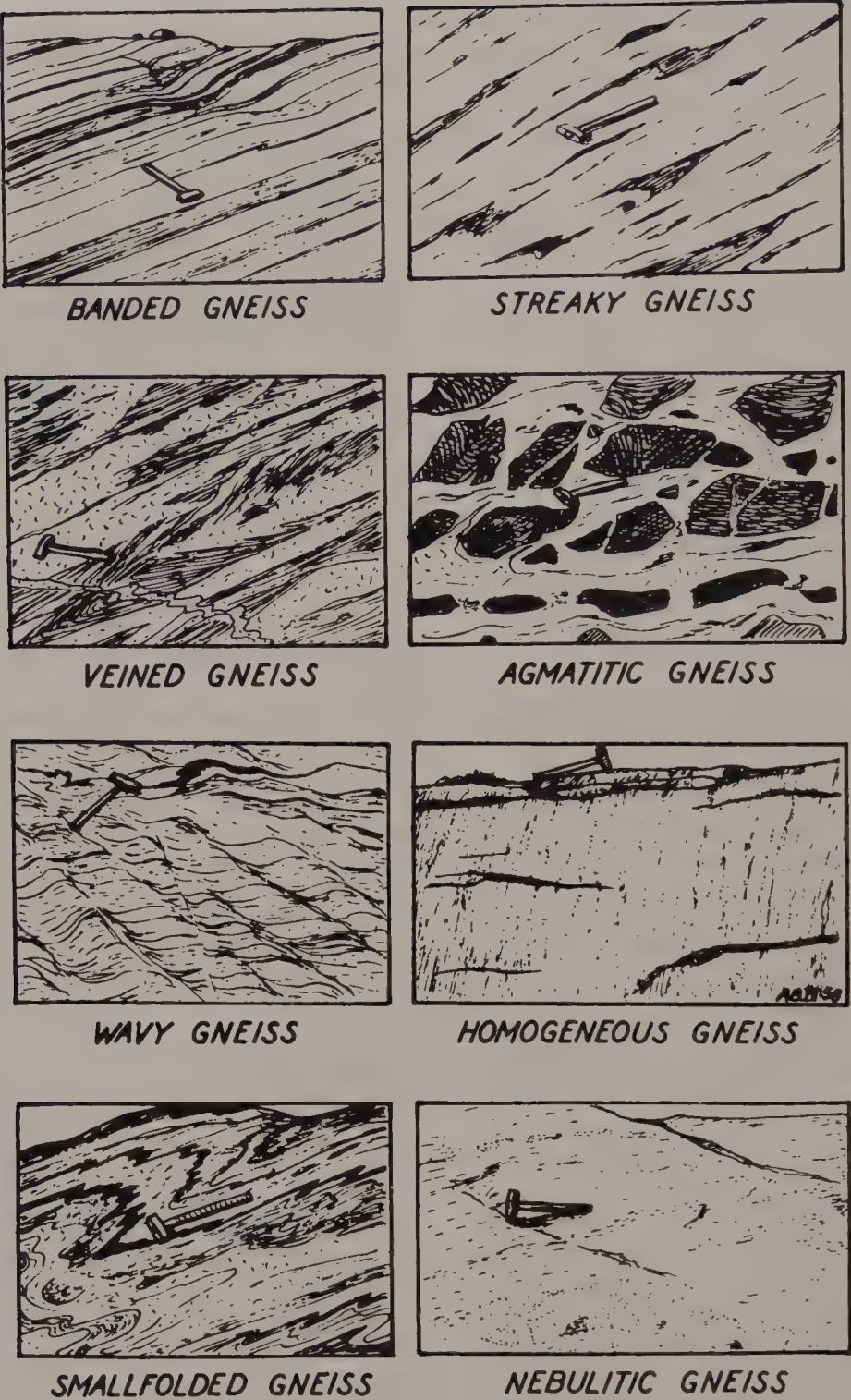


FIGURE 2. Structural classification of gneisses. (From Berthelsen, 1960.)

been used (Fig. 2). Conspicuous textural features characterize some gneisses. These are accordingly termed porphyroblastic gneiss, flecky gneiss, augen gneiss etc. (Mehnert, 1968). Petrogenic varieties based on the supposed origin include orthogneiss (originally igneous), paragneiss (originally sedimentary) composite, injection or mixed gneiss (migmatitic origin) cataclastic gneiss (e.g., stronalite), and mylonite

gneiss (cataclastic origin). Certain gneisses show a combination of varietal attributes, all of which may be indicated by modifiers to the rock name, usually in the following order: structure-mineralogy-petrography-texture-petrogenesis. An example of this is a banded biotite-garnet flecky orthogneiss. Other terms that have been used synonymously with certain gneisses are kinzigite (cordierite-bearing, garnet-biotite

gneiss), khondalite (garnet–sillimanite gneiss), leptyte, leptynite, granulite, and granofels (Katz, 1972).

Origin

Gneisses are formed by a variety of processes. Most are found in high-grade metamorphic terranes and interpreted as the end products of progressive regional metamorphism. Some gneisses can be traced into their less metamorphosed equivalents, e.g., slate → phyllite → schist → gneiss (Fig. 3). The conversion of hydrous, finely foliated, mica-rich rocks such as schists into relatively anhydrous, coarser-grained rocks of quartzofeldspathic composition containing minerals like sillimanite, garnet, cordierite, and pyroxene, results mainly from the breakdown of muscovite and biotite by dehydration under high-grade metamorphic conditions, e.g.,

- (1) muscovite + quartz → K-feldspar + aluminosilicate + H₂O,
- (2) biotite + 3 quartz → K-feldspar + 3 hypersthene + H₂O, and
- (3) biotite + muscovite + quartz → garnet (high *P*) or cordierite (low *P*) + K-feldspar + H₂O.

Since reaction (1) proceeds at moderate temperatures (~600°C) and reaction (2) at much higher temperatures (~800°C), under the normal range of high-grade metamorphism (amphibolite–granulite facies), biotite tends to remain stable while muscovite completely disappears. At the same time that muscovite is being converted into feldspar and aluminosilicate (e.g., sillimanite), the temperature is high

enough to cause grain growth, recrystallization, and coarsening of the texture. Pre-existing biotite-rich foliae may survive to contribute to the layered appearance of the gneiss.

Many quartzofeldspathic gneisses have been derived from crystalline igneous rocks (Myers, 1978) and these orthogneisses may form under a range of igneous and metamorphic conditions. In parts of most granitic bodies the rocks display foliation or lineation that resembles that of a gneiss. Some of these gneisses may be formed under primary igneous conditions. In some instances these gneissic zones are restricted to the margins of the igneous mass, displaying evidence of deformation and cataclasis and may be associated with protoclastic effects at contacts with country rocks. Other gneisses, which show no evidence of cataclasis, may reflect primary processes related to the emplacement of the rock. The foliation in this case is assumed to be parallel to flow surfaces of the magma. A third group of gneisses may have been emplaced as discordant igneous bodies but have subsequently been deformed and metamorphosed along with the country rock, thus assuming conformable structures. These pre-tectonic orthogneisses have much in common with syntectonic types that were intruded at the same time as the regional deformation.

In high-grade terrane, granitization or partial melting (anatexis) may mobilize the quartzofeldspathic constituents relative to the more refractory mafic components to form a variety of gneisses. The gneisses formed by granitization are thought to result from processes by which solid rocks of any composition are transformed into granitic rocks mainly by metasomatism without passing through a mag-

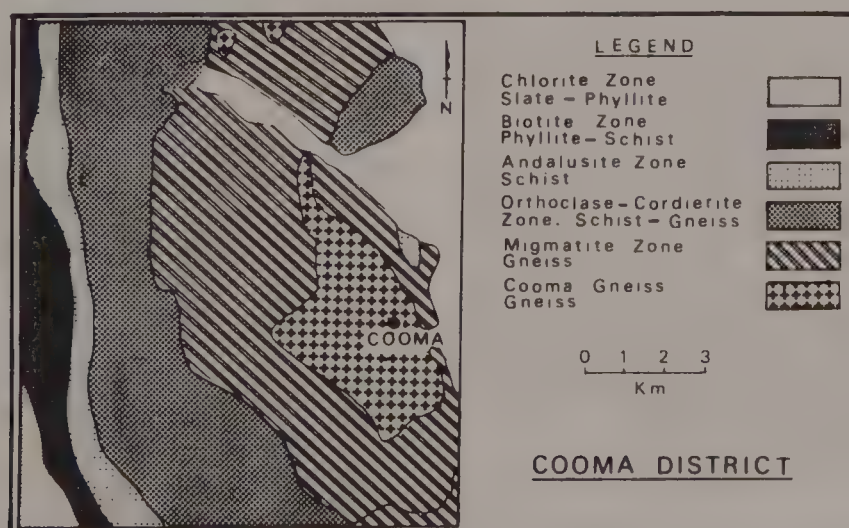


FIGURE 3. Progressive regional metamorphism of metapelites of Cooma, N.S.W., Australia. (From Joplin, 1942.)

matic stage. The gneissic structure in this case may reflect original bedding or structure in the parent rock. Migmatitic gneisses may be formed in marginal zones of igneous intrusions or in high-grade metamorphic terranes by partial melting. The structural relationships between the mobile quartzofeldspathic "leucosome" and the more resistant mafic "melanosome" can be very complex and many varieties of migmatitic gneisses are known (e.g., phlebitic or veined gneiss, stromatitic or layered gneiss, and pygmatic gneiss—Mehnert, 1968; see **Migmatite**).

Many rocks may be converted into gneisses by progressive deformation during metamorphism (Borradaile et al., 1982). At low metamorphic grades the rocks are fractured and a series of closely-spaced discontinuities are formed. With increasing deformation the minerals themselves become strained, are flattened, and are reduced to a fine-grained matrix. The differential mechanical behavior of the granular and platy and prismatic minerals is such that they tend to segregate into alternating fine bands parallel to the movement directions in the rock to form a mylonite gneiss (Fig. 4). Under the most extreme conditions, melting occurs to produce trapshotten gneiss (a gneiss injected with psuedotachylite). Under higher grades of metamorphism, recrystallization accompanies the deformation to produce a series of rocks that range from less-deformed, patchy, lenticular gneisses to highly deformed, parallel, layered gneisses (Myers, 1978).

The exact origins of many gneisses are still not clear, especially in cases where field relations are obscure, original textures and structures have been destroyed, and where original compositions are not diagnostic of

sedimentary or igneous origins. This is prominently shown in those well-layered or banded gneisses that show a regular alternation of light and dark bands reminiscent of bedding. These rocks may reflect original sedimentary or igneous-volcanic layering, structural discontinuities, or even metamorphic differentiation bands (Katz, 1970).

Augen gneiss

Augen gneiss is a general term for gneissose rocks containing phacoidal (lenticular) crystals or aggregates (from German *auge*, eye). The augen may represent uncrushed fragments (porphyroclasts) or porphyroblasts. These rocks are distinguished from ordinary gneisses by their anastomosing or curvilinear anisotropism, formed by mineral laminae sweeping around lens-shaped, phacoidal crystals and aggregates. The term *augen*, originally defined by C. Lapworth in 1885, designates eye-like structures and textures in a variety of gneissose igneous and metamorphic rocks. Most augen gneisses have been dynamically metamorphosed and such rocks have affinities with mortar gneisses, flaser gneisses, mylonite and blastomylonite gneisses, and cataclastic and protoclastic gneisses (see **Cataclastic rocks**). Cataclastic, protoclastic, and mylonite augen gneisses are formed by the progressive deformation of the rock, with the augen representing uncrushed fragments or porphyroclasts. Where recrystallization accompanies the deformation, a sequence of textural changes (mortar gneiss → augen gneiss → flaser gneiss → granoblastic gneiss) has been described (Fig. 5). This is especially well shown in rocks of quartzofeldspathic composition. With increasing deformation, the feldspars are deformed into lens-shaped augen and progressively recrystallized, while the quartz and mafic minerals tend to sweep around the feldspar porphyroclasts.

Not all augen gneisses are formed as a result of dynamic metamorphism of pre-tectonic

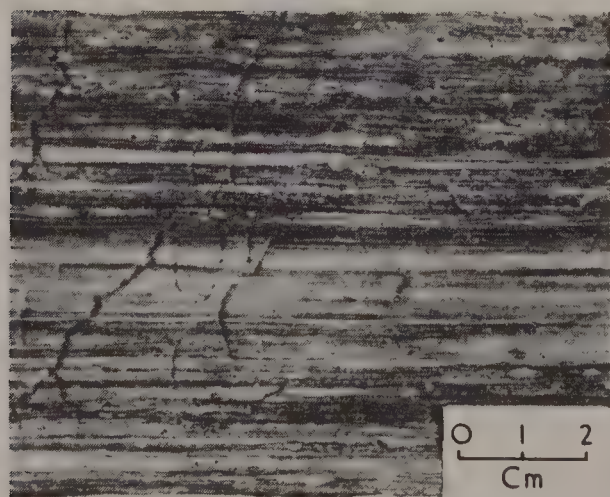


FIGURE 4. Mylonite gneiss; dark layers of mica and chlorite alternate with layers rich in quartz or microcline; Wyangala Dam, N.S.W., Australia.

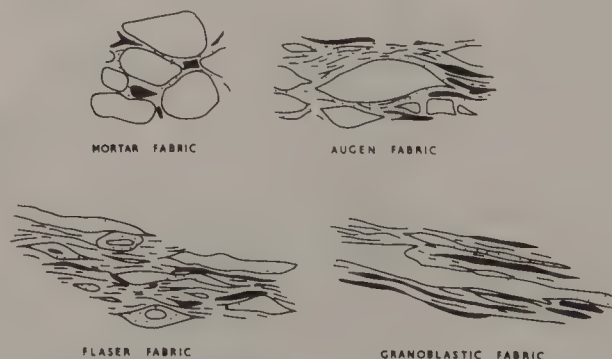


FIGURE 5. Sequence of textural changes accompanying progressive deformation and recrystallization of coarse-grained quartz mangerites; Mount Tremblant Park, Quebec, Canada.

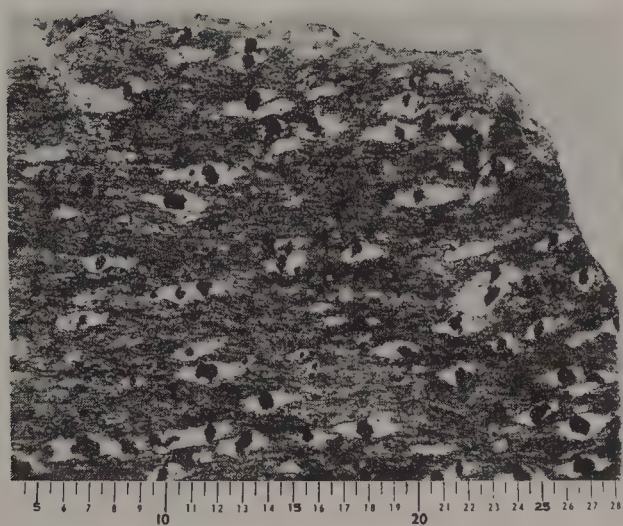


FIGURE 6. Biotite flecky gneiss; dark spots or flecks of biotite contained within a light-colored lenticular augen rim of quartz-feldspar depleted in biotite, set in a biotite gneiss; Broken Hill, N.S.W., Australia.

crystals; some are formed under syntectonic conditions in which crystals grow synchronously with tectonic movements as porphyroblasts. In this case the crystal has not been highly deformed or recrystallized but may assume a lens-shaped form. Other types of augen gneisses are formed by segregation of feldspar (and quartz) crystals and aggregates during migmatization (Mehnert, 1968). Where these augen are closely spaced, the rock is termed a *pearl gneiss*. Flecky gneisses are related to these migmatitic augen gneisses and are formed by segregation of the dark components of the parent rock into flecks, leaving behind a halo depleted in these constituents (Fig. 6).

M. B. KATZ

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Cross-references: *Anatexis*; *Banding in igneous rocks*; *Cataclastic rocks*; *Injection complex*; *Khondalite*; *Metamorphism*; *Migmatite*; *Migmatite—origin of neosome*; *Migmatite—structural relationships*; *Porphyroblast*; *Regional metamorphism*; *Rheomorphism*; *Schist*.

GRANITE

Granite is used as a geologic term in two ways. Strictly, it is used for a plutonic igneous rock of very specific mineralogy; broadly, it is used for a group of plutonic igneous rocks with modal quartz, often referred to as granitic or granitoid.

Granites *sensu stricto* are phaneritic, alkali-rich, silica-oversaturated rocks consisting of one or two feldspars, quartz, and a variety of mafic and opaque phases. Both granites and granitoids are classified by the IUGS recommendations on their feldspar composition and quartz:feldspar ratio (QAP diagram—Fig. 1). Thus, all these rocks contain 20–60% quartz as a fraction of the felsic minerals; this field is subdivided by arbitrary ratios calculated as plagioclase(P)/alkali feldspar(A) + plagioclase(P), where plagioclase is taken as An_{10–100} and alkali feldspar is orthoclase or microcline plus albite (Streckeisen, 1976). Chemical analyses and normative mineralogy are given in Table 1.

A number of rock names entrenched in the literature do not appear on Fig. 1, and the

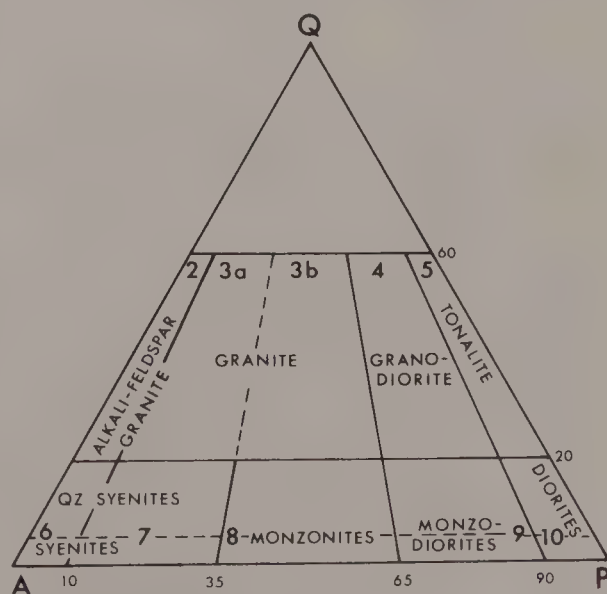


FIGURE 1. IUGS recommended nomenclature for granitoids and related rocks. (After Streckeisen, 1976.)

TABLE 1. Chemical analyses and normative mineralogy of granites and granitoids

	1	2	3	4	5	6
SiO ₂	71.30	68.65	73.86	73.28	71.45	73.83
TiO ₂	0.31	0.54	0.19	0.21	0.34	0.28
Al ₂ O ₃	14.32	14.55	11.71	13.12	13.86	13.04
Fe ₂ O ₃	1.21	1.23	1.75	1.16	0.78	1.88
FeO	1.64	2.70	1.47	1.65	2.35	2.12
MnO	0.05	0.08	0.06	0.06	0.04	0.09
MgO	0.71	1.14	0.17	0.19	0.38	0.40
CaO	1.84	2.68	0.42	0.70	1.08	0.82
Na ₂ O	3.68	3.47	5.16	4.25	2.77	6.17
K ₂ O	4.07	4.00	4.61	4.88	5.89	0.37
P ₂ O ₅	0.12	0.19	0.03	0.03	0.10	0.05
H ₂ O	0.77	0.73	0.37	0.57	0.62	1.09
TOTAL	100.02	99.96	99.80	100.07	99.66	100.33
Norms						
qz	29.06	25.17	28.30	27.99	28.77	34.74
cor	0.92	0.28	—	—	1.42	2.35
or	24.50	23.66	27.24	28.12	34.88	13.34
ab	31.13	29.36	34.58	35.63	23.43	31.40
an	8.04	11.55	—	3.41	4.45	2.11
di	—	—	1.94	1.70	0.10	—
hy	3.37	5.66	0.72	1.58	4.16	8.57
ol	—	—	—	—	—	—
ac	—	—	3.93	—	—	—
ns	—	—	2.02	—	—	—
mt	1.75	1.79	1.30	1.67	1.13	2.78
il	0.58	1.03	0.36	0.39	0.68	0.61
ap	0.28	0.44	0.07	0.23	0.26	1.24

- 1. Average granite (Le Maitre, 1976).
- 2. Average adamellite (Le Maitre, 1976).
- 3. Peralkaline granite, Nigeria (Bowden and Turner, 1974).
- 4. Non-peralkaline granite, Nigeria (Bowden and Turner, 1974).
- 5. Rapakivi granite, Finland (Vorma, 1976).
- 6. Trondhjemite, Ballantrae Complex (ophiolitic), Scotland; (Jelínek et al., 1984).

IUGS committee regard them as obsolete. These include *adamellite* of European geologists, and *quartz monzonite* of American geologists; both correspond to subarea 3b. There are other rocks of petrogenetic importance with well-established names that do not appear in a classification that does not distinguish between sodic and potassic alkali feldspar. The most important of these are *trondhjemite* and *plagiogranite*, where most of the feldspar is albite or oligoclase. These rocks typically have a very low color index and very low K₂O content; they could be regarded as leucogranodiorite or leucotonalite in the QAP scheme.

Mafic minerals do not influence this classification. They can include micas, amphibole, pyroxenes (clino- or ortho-), rare olivine (fayalite), and tourmaline. Accessory phases are usually oxides (magnetite or ilmenite),

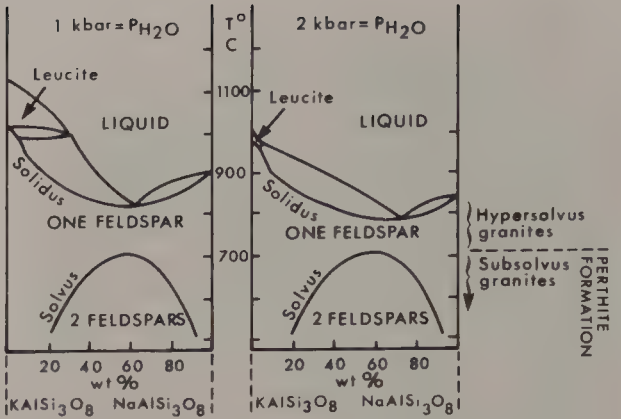


FIGURE 2. The system KAlSi₃O₈-NaAlSi₃O₈-H₂O to illustrate the relationship of feldspar type and textures to granite crystallization. (After Tuttle and Bowen, 1958.)

zircon, sphene, apatite, or rutile, while occasionally monazite, epidote, fluorite, cordierite, aluminosilicates, and topaz may be important.

Mica is usually biotite and/or muscovite, while Li-bearing species like lepidolite and zinnwaldite are known, but rare. The most common amphibole is hornblende, while, in strongly sodic granites, minerals like hastingsite, riebeckite, and richterite are recorded. Likewise with pyroxenes, which are usually sodic forms such as aegirine. The occurrence of olivine, usually precluded by the presence of quartz or hypersthene in the norm, is restricted to anhydrous, Fe-rich, alkali granites (such as those in the British Tertiary Volcanic Province) where an Fe-rich orthopyroxene appears in the norm. Granitic rocks with modal orthopyroxene, usually hypersthene, are members of the charnockite suite (as distinct from charnockite).

Feldspars in granitoids often have a complex paragenesis. Granitic magmas that crystallize at temperatures above the alkali feldspar solvus (hypersolvus granites) contain one feldspar, while those that crystallize at subsolvus temperatures (*subsolvus granites*—the majority) contain two (orthoclase or microcline plus a sodic plagioclase). *Hypersolvus granites* contain perthitic feldspar grains, where the originally homogeneous feldspars have exsolved into a two-feldspar intergrowth (Fig. 2).

Features of fabric

The texture referred to are not restricted to granites *sensu stricto*. Their variety and complexity reflects and provides evidence for the often intricate petrogenesis of these rocks.

Porphyritic texture is widespread. The phenocryst phase (megacrysts when the grains are

>10 cm across) is generally a feldspar, and may be aggregated. The most spectacular manifestation is seen in *rapakivi* (Finnish, “crumbly stone”) texture in which large orthoclase phenocrysts or aggregates are rimmed with oligoclase. Other feldspar phenocrysts often show complex twinning, zonation and evidence of resorption.

Synneusis texture (Greek, “swimming together”) involves the presence of chadocrysts of one or more phases, usually biotite, amphibole, or pyroxene, enclosed in larger grains, usually feldspar.

Graphic, *runic*, or *cuneform* texture is more common in pegmatite than in granite. The quartz of quartz-feldspar intergrowths may look like script, particularly on a cut or polished surface, hence the various synonyms. In some instances the quartz intergrowth patterns resemble hieroglyphs. Phenocrysts or xenoliths alone may show preferred orientation, reflecting the sluggish motion of a viscous crystal mush, or, as the result of deformation, the whole pluton may possess a penetrative, gneissose fabric.

Origin

The origin of granite has proved a controversial subject from the earliest days of modern geology, when A. G. Werner proposed the precipitation of crystalline rocks from a primeval ocean. The “granite controversy” (Read, 1957), at least in part, reflected the frequently equivocal relationship between granitic plutons and their country rocks, especially in high-grade metamorphic terranes where migmatites are developed. One much debated aspect was the significance of feldspar phenocrysts in granitic plutons that apparently matched porphyroblasts of the same mineral in the surrounding aureole rocks and in xenoliths. The debate was further stimulated by the apparently gradational relationship between some granites and granitic gneisses. One (transformist) school of thought contended that many, or even all, granitic rocks resulted from metasomatic granitization of country rock by invading K-rich “ichors”; the other (magmatist) school contended that all granitic rocks had been molten at some stage in their history.

Much of the controversy was resolved by the experimental studies of Tuttle and Bowen (1958) who investigated phase relationships in the synthetic system quartz-orthoclase-albite-water (Q-Or-Ab-H₂O) at various values for P_{H_2O} (Fig. 3a). They found that most actual compositions of granitic rocks are very closely concentrated around the minimum melting composition in this system at P_{H_2O} of 1–2 kb

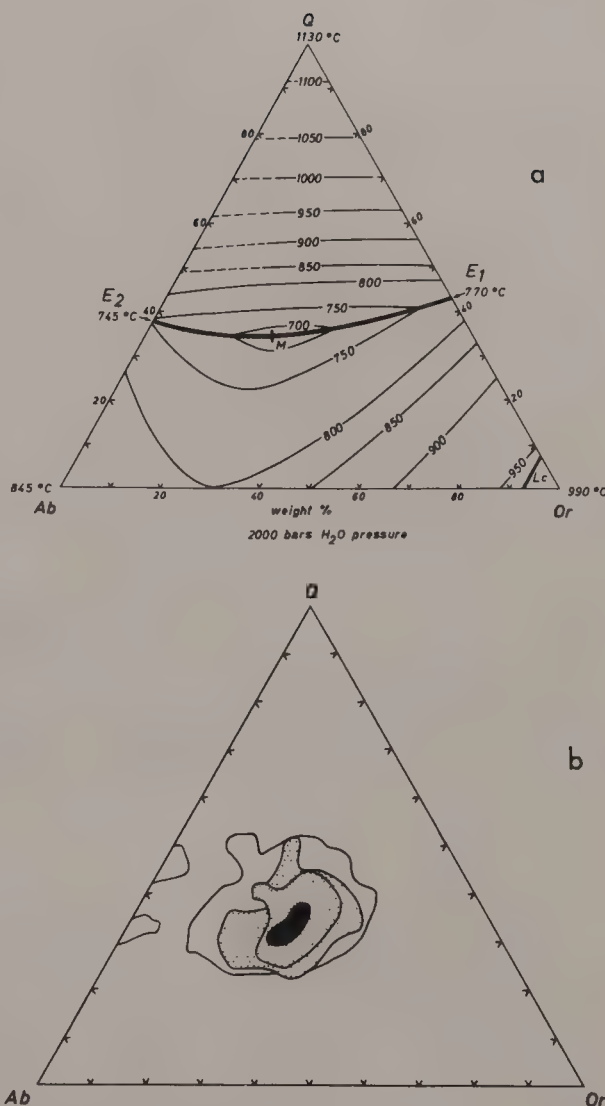


FIGURE 3. (a) Phase relations in the system Q-Ab-Or-H₂O at 2 kb P_{H_2O} ; thick line is the boundary between silica and alkali feldspar fields; M is the minimum melt composition on the cotectic line. (From Winkler, 1967.) (b) Normative compositions of 1190 granitic rocks plotted in the Q-Or-Ab field for comparison with (a); outer line encloses 86% of the sample, coarse stippling covers 75%, and fine stippling and black cover 35%. (From Winkler, 1967.)

(Fig. 3b). The mineralogy of most granites can be explained in terms of crystal-liquid equilibria, and modern consensus supports a magmatic origin for these rocks. Where debate continues it relates to the source of the magma and the nature and extent of liquid-mediated magma-country rock interactions (Taylor, 1988).

The ultimate source of granitic magmas remains a controversial subject, and has been so since Bowen (1928) demonstrated the possibility of producing granitic liquids by the fractionation of a basaltic melt. Green and Ringwood (1968) demonstrated that acidic melts could be produced in large quantities by the partial

TABLE 2. Petrographical, mineralogical, and metallogenic criteria for granite types

M-type	I-type (Cordilleran)	I-type (Caledonian)	S-type	A-type
Plagiogranite subordinate to gabbro	Tonalite to granite with tonalite predominant; minor gabbro	Granodiorite to granite; minor gabbro and diorite	Granites—all high SiO ₂ over narrow range; Li-, B-, and F-rich varieties	Granite and alkali feldspar granite, with syenite and foidal rocks
Hornblende and biotite; pyroxene, magnetite	Hornblende and biotite; magnetite, sphene	Biotite; ilmenite, magnetite	Muscovite, red biotite; ilmenite, monazite, garnet, cordierite	Green biotite, alkali amphiboles and pyroxenes; astrophyllite
K-feldspar is interstitial and micrographic	K-feldspar is pink, interstitial and xenomorphic	K-feldspar is pink, interstitial and invasive	K-feldspar megacrysts (white) often with protracted history	Perthites; rapakivi textures
Basic igneous xenoliths	Dioritic xenoliths, possibly resitite	Mixed xenoliths	Metasedimentary xenoliths dominant	Cognate xenoliths and basic magma blebs
Mol. Al/(Na + K + Ca/2) <1.0	<1.05	ca. 1.0	>1.05	Often peralkaline, relatively F-rich
Initial ⁸⁷ Sr/ ⁸⁶ Sr <0.704	<0.706	<0.709 >0.705	>0.708	0.703–0.712 (considerable range)
Porphyry Cu and Au mineralization	Porphyry Cu and Mo mineralization	Rare mineralization	Sn and W greisen and veins, U–Th, and Li-pegmatites	Sn–W veins, Ta–Nb, fluorite, rare metal and Li-pegmatites
Ophiolites and island arcs	Cordilleran ocean-continent margins	“Andean” and late stages of collision belts	“Hercynian”-type margins	Continental rifts and hot spots

Source: modified from Pitcher (1983).

melting of hydrous basalt. Both mechanisms require that granite is the end product of a process that also generates or involves much larger volumes of more mafic magma. Thus, granites should form volumetrically small fractions of basaltic and, or, andesitic igneous provinces, a relationship only rarely seen. Granitoids do occur in association with these more mafic rocks, but usually form the largest component of the assemblage. Furthermore, granitic rocks, as plutons and gneisses, comprise the bulk of the continental crust.

Isotopic studies, mainly of the Sr, Pb, and Nd systems, have done much to clarify the question of origin, in so far as mantle or crustal melting are considered as alternative sources. Rocks derived directly by partial melting of the mantle (e.g., basalt) have low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.702–0.706), whereas those in the crust, preferentially enriched in radiogenic ^{87}Rb (which decays to ^{87}Sr), have a higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (0.730 average). These ratios are not significantly altered by partial melting or differentiation, so that granites produced by melting of a subducting oceanic slab would have a low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, while melts generated in the crust would have a higher initial ratio.

The study of Pb, U–Pb, and Sm–Nd systematics approaches the problem rather differently. These systems are used in conjunction with radiometric dating to assess the length of time between the final emplacement and cooling of a pluton and the time at which its constituent components first separated from the mantle. Such combined isotopic and geochemical studies of individual granite plutons indicate that some components have long crustal residence histories, while others are close to mantle compositions: many granite magmas have both mantle and crustal components (see various papers in Atherton and Tarney, 1979).

Occurrence

Granitic plutons occur in several distinct tectonic settings. Most occur in orogenic belts, and in the deeply eroded levels of older orogenic belts granite and granitic gneiss may form most of the rock at outcrop. This reflects the relationship between the partial melting of subducted oceanic crust and overlying continental crust at destructive plate margins. Away from such continental margins, granitic rocks form a minor component in the stratiform layered basic intrusions of stable cratonic areas, major components in the ring complexes of rift valleys and flood basalt provinces, and an unknown percentage of layer 2 of the oceanic crust. Some gross chemical and petrologic

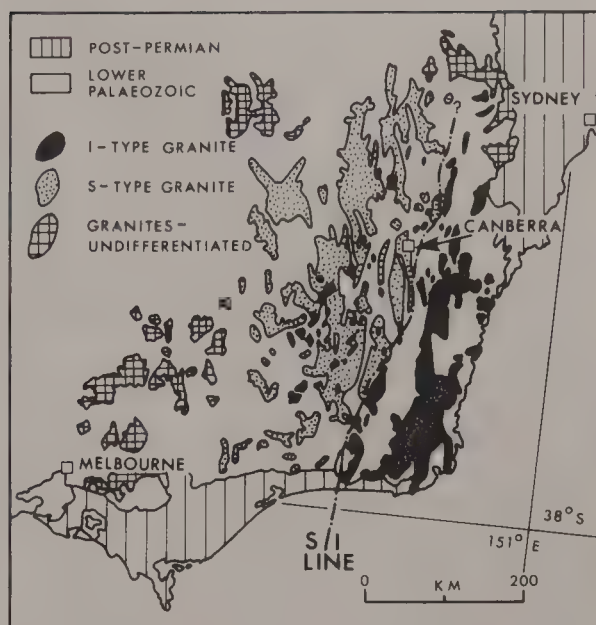


FIGURE 4. Granitoids of the Lachlan Fold Belt, SE Australia, showing the distinction between an area of exclusively I-type granites and one of predominantly S-types with subordinate I-types separated by the so-called S–I line. (After White and Chappell, 1977.)

differences are immediately obvious, in that most intracratonic, anorogenic granitoids are strongly alkalic, both potassic and sodic; the granitic components of ophiolites, and by implication oceanic crust, are plagiogranites and trondhjemite.

Chappell and White (1974) and White and Chappell (1977) suggested that granitoids in an orogenic setting could be subdivided on the basis of a set of mineralogical, chemical, and isotopic criteria into S-type and I-type. This is considered to reflect whether the magma was derived by the melting of a sedimentary or igneous source (White and Chappell, 1988; Chappell and Stephens, 1988). The criteria for subdivision are set out in Table 2; for the granitoids of the Lachlan Fold Belt, SE Australia, they defined two distinct provinces (Fig. 4). Elsewhere, these criteria do not provide such clear-cut subdivisions. Pitcher (1983) has extended this scheme and proposed a correlation between types and tectonic setting (Table 2), by adding M- and A-types (“mafic-related” and “alkali-rich”), and subdividing the I-type. This scheme suggests not only a correlation of granite typology with tectonic setting, but also with metallogenic provinces.

Other terms

Alaskite—the equivalent of alkali granite; commonly used in the U.S.A. for a granite with only a few percent of dark minerals.

Aplogranite—a leucocratic granite with granitic texture and composed essentially of alkali feldspar and quartz.

Binary granite—a synonym of two-mica granite.

Black granite—a commercial term for polished diorite, dolerite or gabbro that is dark gray or black.

Ekerite—a granite that contains arfvedsonite as an essential component and passes marginally into ekerite porphyry (Eker, Norway).

Evisite—alkali granite—alkali syenite of the Evisa area, Corsica.

Fasibitikite—a riebeckite—acmite granite that also contains eucolite and zircon.

Granitite—(1) a granite that contains biotite but no muscovite or other ferromagnesian mineral; (2) a granite rich in oligoclase.

Graphic granite—a granite with graphic texture.

Haplophyre—a granite characterized by large quartz and feldspar grains in a mortar structure.

Kalialaskate—an alaskite that lacks albite.

Kammgranite—a porphyritic hornblende granite.

Karlsteinite—a microcline-bearing alkali granite (Karlstein, Austria).

Kristianite—a local name for biotite granite of the Oslo (Kristiania) region, Norway.

Luxullianite (luxulyanite)—a variety of tourmalinized granite (Luxulyan, England).

Orbicular granite—see **Orbicular rocks**.

Pladorite—a hornblende granite or a mica-hornblende granite.

Protogine (protogene)—a granitic rock with gneissose structure, containing sericite, chlorite, epidote and garnet and showing evidence of recrystallization under stress after consolidation.

Rapakivi granite, pyterlite, wiborgite—see **Rapakivi texture**.

Rockallite—a quartz—albite—microcline granite (Rockall, N Atlantic Ocean).

Runite—a synonym of graphic granite.

Tsingtauite—a porphyritic granite with phenocrysts of micropertthite and sodic plagioclase (Tsingtan, Korea).

Ultrametagranite—a granite formed as a result of extremely high-grade metamorphism possibly involving partial melting.

Yentinite—an obsolete term for a biotite-bearing granitic rock in which the quartz was originally misidentified as scapolite (Yenta River, Alaska, U.S.A.).

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Cross-references: *Batholith*; *British Tertiary Volcanic Province*; *Charnockite suite*; *Diapirism*; *Experimental petrology*; *Granites and mineralization*; *Orbicular rocks*; *Pegmatite*; *Rapakivi texture*; *Textures of igneous rocks*.

GRANITES AND MINERALIZATION

The relationship between granitic rocks and mineralization is governed by the complex interactions of a crystallizing and cooling pluton, circulating fluids and country rock. An empirical relationship between granites,

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hydrothermal fluids, and mineralization has been recognized since the days of Agricola, but the detailed nature of the relationship remains a source of controversy. Particularly controversial are both the source of mineralizing fluids (juvenile magmatic or recycled meteoric water) and the source of the metals.

The end stages of crystallization of a granitic magma often produce fluids rich in water and incompatible elements. They crystallize as pegmatites and aplites both within the pluton and in the bordering country rock. The act of intrusion produces a local thermal anomaly, and even after the magmatic products have finally solidified a granitic pluton remains a focus for fluid circulation, perhaps for millions of years. A certain amount of mixing of juvenile and recycled meteoric water is inevitable, and fluid exchange between pluton and country rock is a two-way process, both before and after crystallization. The form of the mineralization associated with granites is determined by whether such fluid exchange affects either granite, or host, or both, whether the effects are pervasive or structurally channeled, and at which stage they occur in the history of a pluton. The result is a spectrum of ore deposit types, within which four broad categories can be distinguished.

Types of mineralization associated with granites

Pegmatites and aplites. These represent the end-stage magmatic fluids enriched in incompatible elements like Li, Be, K, B, F, Cl, Nb, Ta, Zr, Sn, W, Cs, Ba, P, U, Th, and rare-earth elements (REE). Simple pegmatites and aplites are quartz and K-feldspar rocks, with or without mica, but complex pegmatites, in which F, particularly, appears to have been an important fluid component, contain suites of minerals, commonly as many as ten, or more. Such pegmatites crystallize at temperatures ranging from 900°C to as low as 500°C. These deposits are an important source of Cs, Be, Li, Nb, Ta, mica, and vermiculite, and a locally significant source of Sn, W, U, and Th (see **Pegmatite; Pegmatitic ore deposits**).

Pneumatolytic and metasomatic deposits. These represent a pervasive and bulk alteration of granite, country rock, or both, by either juvenile or recycled meteoric water carrying a variety of complexing agents (F^- , Cl^- , BO_3^{3-} , PO_4^{3-} , HCO_3^- , HS^- , SO_4^{2-}). The names used for the various manifestations are numerous, and reflect the primary mineralogy of the end product. Thus, a granite in which K-feldspar is replaced by fluorine-rich white mica is termed a *greisen*; a metamorphosed limestone showing

progressive silicification and a variety of mineralizations is termed *skarn*, or *tactite* if Fe-rich. Other rock-types are referred to by the process considered responsible for their alteration, usually reflecting the principal new mineral phase, e.g., the products of tourmalinization, albitization, and kaolinization.

Ore minerals in such rocks are usually finely disseminated throughout the host, although local, massive ore segregations are not uncommon. In the case of cassiterite (SnO_2) and wolframite ($(Fe, Mn)WO_4$) in greisen, the ore grade may be quite low (<1%), but the enormous size of some such deposits makes them economically viable. In other cases the bulk of the altered rock is the desired material: in the case of the "china clay" deposits of SW England, up to 50% of the rock consists of kaolinite.

Pneumatolytic deposits, particularly greisen, are a major source of Sn and W, and an important source of Sb, As, Bi, Nb, and Ta. Pneumatolytic kaolinite deposits are the most important source of china clay.

Skarns often have extremely complex mineralogy and contain polymetallic ore deposits in a gangue consisting of many silicate, carbonate, and oxide minerals. The deposits are small but often very rich, occasionally in rare metals or minerals. They form important sources of Ag, B, Be, Bi, Cu, Fe, Mn, Sb, Sn, W, and Zn.

Porphyries and stockworks. These consist of meshworks of veins in a wholly or partially altered host that may be both the pluton or its host-rock. The term "porphyry" in this sense is confusing, referring to the fact that the first major Cu deposits of this type to be exploited, were associated with porphyritic diorite intrusions; subsequently exploited deposits of the same type are not necessarily related to "porphyry" as such. There is, however, a spatial relationship with calc-alkaline plutons, and especially with the upper parts of epizonal plutons in the subvolcanic zone.

The overall form of "porphyry" deposits appears to reflect a water convection cell involved in the cooling of a pluton. The process appears to have two or three distinct phases involving (1) initial K-metasomatism of both the pluton and aureole, followed by (2) a superimposed zoned alteration consisting of an inner phyllic zone (white mica or sericite), and outer argillic zone (clays) and an intermediate propylitic zone (chlorite) (Fig. 1). In some deposits, a third phase involves supergene enrichment, superimposed on everything developed earlier (Fig. 1). Mineralization occurs either as a vein-mesh ("ore-shell") around the pluton contact, or dispersed through the margins of the pluton and the surrounding

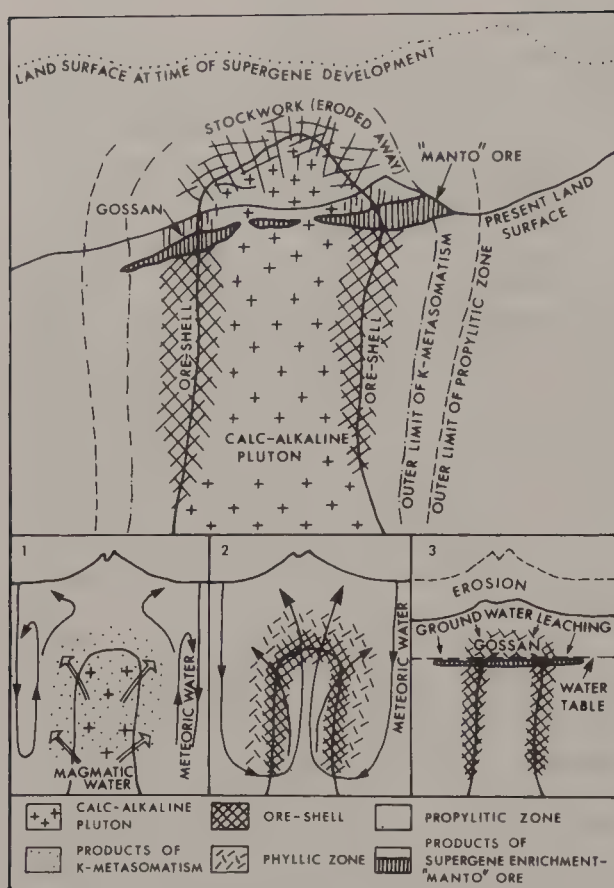


FIGURE 1. Schematic section through a porphyry copper deposit showing the relationship of the alteration zones and the mineralization to the pluton. Inset shows a 3-stage model illustrating the evolution of a water-circulation system during (1) early cooling history, (2) late cooling history, and (3) supergene enrichment phases of porphyry copper deposit genesis.

propylitic zone. Supergene enrichment produces an irregular sheet or "manto" (Spanish, "cloak") that cuts across the earlier zoning (Fig. 1).

The fluid temperature varies from 600°C in the potassic phase, to 300°C in the later and outer alteration zones. Fluid compositions vary considerably but O-H isotope studies and fluid inclusion analysis suggest a magmatic source for the brines responsible for the K-metasomatism, and a meteoric source for those producing the zoned alteration (Fig. 1).

Stockwork mineralization has many features in common with "porphyries," but appears to be confined to the roof zones of granitic plutons. The fine vein networks affect both pluton and country rocks. They do not comprise a homogeneous genetic group like the "porphyries," but their formation appears to have involved both juvenile and meteoric waters, and both chloride- and fluoride-rich brines.

Both types produce low-grade ore-bodies of

great size, usually worked on a vast, open-cast basis. The porphyry copper deposits of the Cordillera in N and S America are the best known examples, e.g., Chuquibambilla and El Salvador in Chile. They are a major source of Cu, and an important source of Au, Ag, Sb, and Te. A second important group of porphyries are the most important source of Mo with W, Bi, Te, and Re as important by-products, e.g., Climax, Colorado (U.S.A.). The Sn-W-Ag-Pb deposits of Bolivia (e.g., Potosi) may represent a third class of Sn porphyries, and the Sb-W porphyries of SE China may be a fourth. Stockwork deposits are important sources of Cu, Au, and Ag (e.g., Bingham Canyon and Butte, U.S.A.), Sn, W, and Sb (e.g., Indonesia and China).

Epigenetic veins. Deposits of this type associated with granitic rocks have been known and worked since antiquity, e.g., Cornwall, SW England, Freiberg, E Germany, and Bohemia, Czechoslovakia. The vein arrays follow fracture patterns that may relate to the stresses generated around the intruding pluton. Deposits within these veins often show a thermal zonation of metals and/or mineral associations suggesting that the granitic pluton was the source of heat, if not the fluids and metals as well. The highest-temperature veins often carry silicate minerals (e.g., topaz, tourmaline, zinnwaldite) and are dominated by oxide ore minerals. The wallrock alteration includes greisen. Progressively lower-temperature assemblages occur farther away from the pluton, dominated by sulfides then sulfosalts and sulfates. Wallrock alteration may include tourmalinization, greisenization, and kaolinization.

The vein arrays over the Cornubian batholith of SW England are one of the best-studied examples of this type of mineralization (Fig. 2). The zonation there is particularly well developed (Fig. 3), and the complex relationships between the different types of mineralization has been elucidated. Similar developments occur throughout the Hercynian belt of western and central Europe and the common genetic association of mineral deposits and batholiths has led to the development of terms to describe spatial relationships.

Acrobatholithic—refers to a mineral deposit occurring in or near an exposed batholith dome or the stage of batholith erosion in which that area is exposed.

Cryptobatholithic—refers to a mineral deposit occurring in the roof rocks of an unexposed batholith.

Embatholithic—refers to a mineral deposit occurring in a batholith in which exposure of the

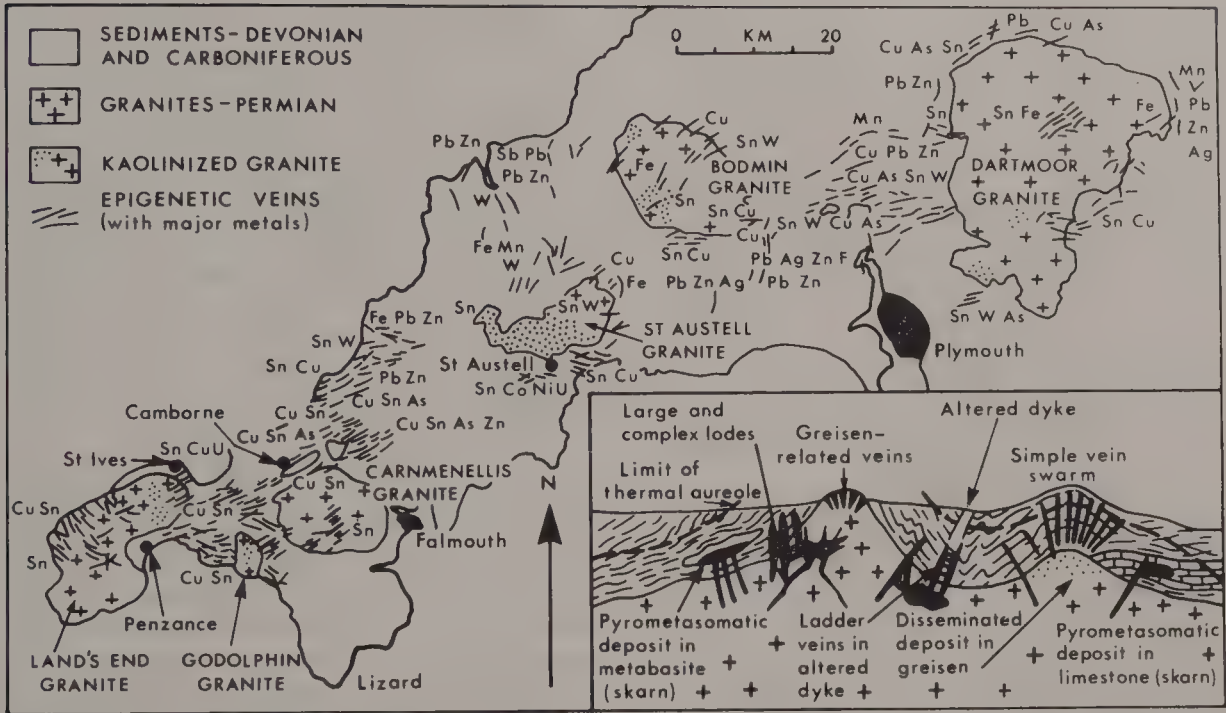


FIGURE 2. Distribution of epigenetic vein deposits in the roof zone of the Cornubian batholith, SW England. Inset illustrates the relationship of different types of mineralization to the granite batholith.

batholith and of the country rock is about equal.

Endobatholithic—refers to a mineral deposit occurring in or near a roof pendant in a batholith.

Epibatholithic—refers to a mineral deposit occurring in the peripheral area of a batholith.

Hypobatholithic—refers to a mineral deposit occurring in the deeply eroded region of a batholith.

In the past, epigenetic vein deposits formed important sources of Pb, Zn, Cu, Fe, Bi, Sb, As, and Mo, but at present only Sn, W, Au, Ag, Hg, and U are considered viable in these rich, but usually small deposits.

Uranium and granites

Many granites contain disseminated U and Th, usually concentrated in the accessory

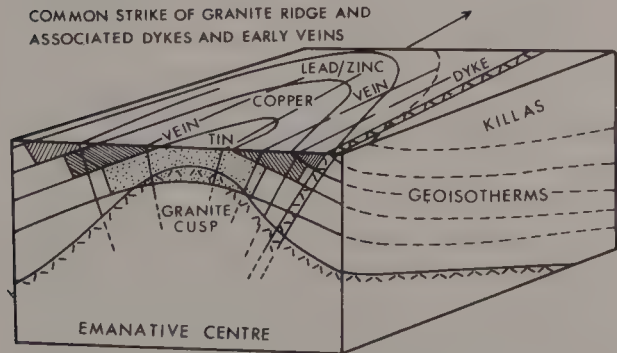


FIGURE 3. Schematic relationship of veins, zonation and geoisotherms to a granite cusp (cupola).

minerals zircon, sphene, monazite, allanite, and apatite. Pneumatolytic metasomatism may concentrate these elements, particularly the process known as albitization, which involves the desilicification and enrichment in alkalis of a granite host.

Uranium concentrations may be up to ore grade in most types of granite-related mineralization, especially pegmatites (e.g., Rossing, Namibia) and hydrothermal epigenetic veins (e.g., Cornwall, SW England and Jachymov—previously Joachimstal—Czechoslovakia). The most distinctive group of deposits, which do not readily fall into the above classification, are those associated with early–middle Proterozoic palaeosols developed over Archaean or early Proterozoic granitic rocks. They formed during the deep weathering beneath an unconformity, preferentially concentrating U with Cu, V, Pb, and Zn, usually as phosphates in particular soil horizons. The huge deposits at Rum Jungle and Alligator River in N Australia are of this type, and smaller examples are known elsewhere, e.g., Paukkajanvaara, Finland.

Placers derived from granitic terrane

Many of the different types of deposit outlined above are subeconomic, and many accessory minerals in granite and pegmatite are too disseminated in their hosts to be considered for exploitation. Such minerals include magnetite, rutile, ilmenite, and zircon, which like cassiterite, wolframite, tantalite, columbite, and

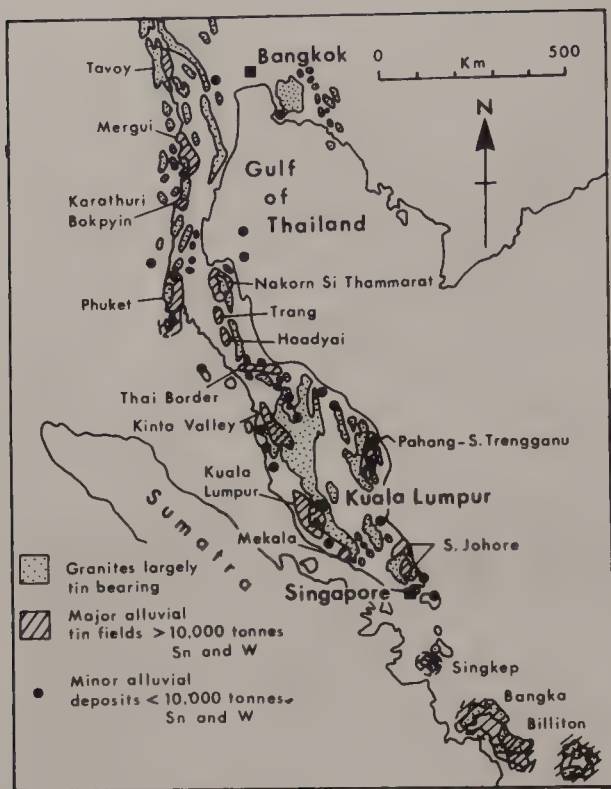


FIGURE 4. Distribution of Sn-W placer deposits to Sn-W-bearing granites in the tin belt of SE Asia.

many oxides of rare-earth elements are both resistant to abrasion and dense. These tend to accumulate during sedimentary winnowing in river gravels and beach sands, where they give rise to important placer deposits. Au is an obvious example, but Sn and W placers are important in SE Asia (Fig. 4). Sn-Nb-Ta placers are exploited in Nigeria, and Ti, REE, Nb, and Ta as well as refractory raw materials are extracted from black sand beaches in E Australia, S India and Brazil.

Tectonic setting of granites and metallogenic provinces

Several attempts have been made to classify granites according to their plate-tectonic setting and petrogenetic characteristics (see **Granite; Igneous rocks—tectonic setting**). Pitcher (1983) has produced a classification that bears some comparison with the metallogenic scheme devised by the Russian workers Tausen and Kozlov (see Levinson, 1974). The most striking correlation between granitoid-related mineralization and plate-tectonic situation is seen in the case of the subduction-related I-type granites and porphyry ore deposits (Cu, Mo, and Sn types). All the Mesozoic and Cenozoic examples of this type are related to either known subduction zones (in the circum-Pacific belt) or subduction zones active prior to continent-continent collision (e.g., Balkans to



FIGURE 5. Worldwide distribution of granite-related metallogenic provinces.

Iran—Fig. 5). Most Sn-W deposits are associated with orogenic S-type granites, e.g., SW England, SE Asia, and SE Australia), and the concentration of Sn appears to involve melting of crustal rocks—invoked where S-type granites overlie subductions zones that descend beneath older continental crust. In contrast, the Nigerian Sn province is associated with anorogenic, rift-related, alkali granites and syenites (A-type of Pitcher, 1983) (see Fig. 5).

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Cross-references: *Aureole; Batholith; Granite; Greisen; Hydrothermal processes; Igneous rocks—tectonic setting; Metasomatism; Pegmatite; Pegmatitic ore deposits; Pneumatolysis; Skarn.*

GRANITES AND VOLCANISM

In many parts of the world, granitic rocks, which include granite, adamellite (quartz monzonite), granodiorite and quartz diorite (tonalite), have been exposed by deep erosion, with any associated volcanic rocks removed. However, in many areas granites and volcanic rocks are associated, and probably crystallized from common parent magmas. One example is

the late Cretaceous Boulder batholith of Montana, U.S.A. (Hamilton and Myers, 1962) and the coeval Elkhorn Mountains volcanic assemblage that oversteps the margin of the batholith on to the country rock. Likewise, the Idaho batholith has associated acidic flows and welded tuffs that possibly acted as its roof (Hamilton and Myers, 1967), while in the White Mountains of New Hampshire ca. 3500 m of flows, tuffs, and breccias are considered to be comagmatic with plutonic rocks including adamellite (quartz monzonite), granite, diorite, and gabbro. There, fracturing and ring dyke formation followed lava extrusion. A generally comparable development is shown in N Nigeria where the extrusion of large volumes of rhyolite (with welded tuff, etc.) during an initial volcanic stage, was followed by emplacement of granitic intrusions presumed to have been from the same magma source (Jacobson et al., 1958).

The association of ring dykes with coeval granitic plutonism and volcanism in areas where cauldron subsidence was operative seems to be worldwide (see **Ring dyke**). One region where the petrogenetic relationship has been studied in detail is in the Peruvian Andes (e.g., Thorpe and Francis, 1979). In New Mexico, U.S.A., where volcanic rocks are common at the surface, geophysical techniques have been used to demonstrate the structural features controlling the shapes of granitic plutons at depth, e.g., in the Valles Caldera (An Keny et al., 1986). Geochemical data from silicic ash flow deposits have been used to reconstruct the volcanic stage of pluton development (Macdonald and Smith, 1988).

C. K. SEYFERT

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- Cross-references: *Batholith*; *Calderas and volcano-tectonic depressions*; *Explosive volcanic eruptions—classification*; *Granite*; *Ring dyke*, *Volcanoes and volcanism*.

GRANODIORITE

The term granodiorite was first proposed by G. F. Becker and published in 1893 by W. Lindgren. In its original definition, the term covered all plutonic rocks intermediate between granite and diorite, i.e., quartz-bearing rocks containing more plagioclase than K-feldspar. When W. C. Brögger introduced in 1895 the group of (quartz-) monzonites, Lindgren restricted the term granodiorite to rocks with a plagioclase:K-feldspar ratio >2 . Johannsen (1932) defined granodiorite as a plutonic rock with a plagioclase:K-feldspar ratio between 95:5 and 50:50, an absolute K-feldspar content of at least 5%, a quartz content $>5\%$ and an amount of mafic minerals; between 5 and 50%.

According to the recommendation of the Igneous Rock Committee of the International Geological Congress in its meeting at Montreal (1972), *granodiorites* are igneous or igneous-looking plutonic rocks composed of 20–60% quartz, calculated on the quartz–feldspar fraction only, with a plagioclase:K-feldspar ratio between 65:35 and 90:10, and containing $<90\%$ mafic minerals; the average plagioclase should carry $<50\%$ anorthite (see Streckeisen, 1976, and **Plutonic rocks—classification and nomenclature**).

Macroscopic features. Granodiorites are generally grayish-white, granular rocks, and because of the lower K-feldspar content, reddish colors, even in weathered rocks, are much less common than in granites. The typical texture is hypidiomorphic-granular. The mafic minerals are the most idiomorphic (euhedral).

K-feldspar, if present in larger amounts is xenomorphic to hypidiomorphic (anhedral to subhedral); it occurs in small proportions, together with quartz, mainly as interstitial filling. Most granodiorites are more or less equigranular, but porphyritic textures are not rare; orbicular textures occur occasionally. The specific gravity of bulk samples is 2.68 ± 0.12 .

TABLE 1. Modal composition of granodiorites

Mineral	Range (%)	Average (%)
Quartz	(14.0) 16.0–43.2 (46.2)	26.9
K-feldspar	(4.6) 4.6–23.2 (25.2)	14.4
Plagioclase	(23.0) 30.2–59.6 (59.6)	44.9
Biotite	(0) 0–17.8 (28.7)	8.3
Hornblende	(0) 0–9.8 (21.6)	2.4
Pyroxene	0–11.2	trace
Muscovite	0–17.6	0.7
Sphene	(0.1) 0.2–1.0 (2.1)	0.7
Apatite	(0.1) 0.1–1.5 (2.2)	0.4
Zircon	(0) 0–0.3 (0.3)	0.1
Epidote	(0) 0–4 (11.9)	~0.5
Opaque ores	(0.1) 0.2–1.0 (6.8)	0.7

Figures in parentheses are extreme values; the ranges not in parentheses are chosen to include at least 95% of all values reported.

Mineralogical and chemical composition.

The modal mineralogical composition of granodiorites is summarized in Table 1. The color index varies from <5 (leuco-granodiorite, 8%) to 5–30 (meso-granodiorite, 91%) and 30–50 (mela-granodiorite, 1%). The biotite:hornblende ratio decreases as the proportion of quartz decreases. The chemical composition of granodiorites varies within wide limits (Table 2), with that for the average igneous rock (calculated from Ahrens, 1965) falling within the limits towards the more basic end of the range.

Varieties. *Alsbachite* is a porphyritic granodiorite or aplite, often with accessory garnet and mica.

Aplodiorite is a light-colored biotite granodiorite containing little or no hornblende.

Banatite is an ill-defined term, including both quartz-poor hypersthene granodiorite and orthoclase-bearing tonalite to quartz diorite. The rock from the type locality plots chemically into the field of basic granodiorites.

Farsundite. The rock from the type locality Vest Agder in Norway is a pyroxene granite in which >65% of the total feldspar is K-feldspar. The name has been used also for hypersthene-hornblende adamellites and biotite-hornblende-quartz monzonites and granodiorites. It has also been used with reference to rocks associated with some anorthosite masses (see **Anorthosite**).

Granogabbro is a rare granodiorite with gabbroic affinities; it contains >50% labradorite.

Opdalite is a basic granodiorite transitional to quartz monzodiorite and tonalite. The type rock from the Opdal Inlet in Norway is a basic pyroxene granodiorite with low Niggli *al* and high *fm* values. It should be termed opdalite, opdalo-enderbite, or hypersthene granodiorite.

Trondhjemite. Although some granodiorite belongs to the trondhjemitic magma type of Niggli, the term trondhjemite has been used to refer to leucocratic granodiorite or quartz-rich tonalite, with albite-oligoclase as the sole feldspar. Mafic minerals constitute <10%, and the rock is K-poor. *Plagiogranite* is also used (see **Granite**).

Effusive equivalents of granodiorites are widespread. They have been described variously as dacites (see **Andesite and dacite**), quartz latites, or rhyodacites.

Occurrence. Granodiorites form large areas within Precambrian massifs (these types are frequently gneissose) and plutons of various size

TABLE 2. Chemical composition of granodiorites (wt.%)

	Range ^a	Average ^a	Average igneous rock ^b
SiO ₂	(55.88) 61.20–74.82 (77.58)	68.79	60.35
TiO ₂	(0.02) 0.02–1.00 (1.64)	0.47	0.95
Al ₂ O ₃	(11.13) 13.05–16.98 (20.30)	15.22	15.50
Fe ₂ O ₃	(0.0) 0.0–2.23 (4.74)	1.26	3.32
FeO	(0.0) 0.17–3.86 (7.56)	2.03	4.24
MnO	(0.0) 0.00–0.18 (1.96)	0.07	0.13
MgO	(0.0) 0.06–2.86 (4.38)	1.30	3.24
CaO	(0.45) 1.02–6.30 (6.80)	3.12	5.88
Na ₂ O	(2.00) 2.51–4.97 (5.99)	3.69	3.24
K ₂ O	(1.15) 1.48–4.46 (5.60)	3.06	2.52
P ₂ O ₅	(0.0) 0.00–0.49 (0.63)	0.16	0.25
H ₂ O ⁺	(0.10) 0.15–1.17 (1.96)	0.67	—

^a Figures in parentheses are extreme values; the ranges not in parentheses are chosen to include at least 95% of all values reported.

^b Calculated from Ahrens (1965).

within orogenic belts. These are frequently exposed at a medium to low crustal level. Some plutons consist of an almost uniform granodioritic rock. Elsewhere, the granodiorite forms irregular areas within plutons of different composition or occurs as a core or border facies of mostly more acidic plutons. Dykes of granodiorite (or microgranodiorite) are common; aplites and pegmatites exist but are rare.

The quantitative importance of granodiorities within the spectrum of all plutonic rocks is difficult to estimate. They are certainly a very common if not the most common plutonic rock-type. Within the South California batholith (2,200 km²), granodiorites account for 34%, but are quantitatively subordinate to tonalites (55–60%); gabbros account for only 14% and granites for 2.5%. Within the very large Cordilleran superprovince of western N America (California, Oregon, Washington, Idaho, Montana, Nevada), the classical area of granodiorites, these rocks form 19% of the total, less than quartz diorites (33%) and quartz monzonites (28%), but more than gabbros and ultrabasic rocks (11%), granites (7%), diorites (2%), and syenites (0.1%). In the Precambrian basement of Sinai, Israel (12,000 km²), granodiorites form about 1% only, the dominant rocks being granite, quartz monzonite, and diorite. Within the Caledonian belt of Scandinavia, granodiorites are subordinate to diorites, norites, and mainly tonalites. It is probably true that rocks most frequently accompanying granodiorites are more basic ones—diorites and quartz diorites—while in a truly granitic environment granodiorites are rather rare. They are conspicuously absent from provinces of alkaline affinities.

Granodiorites are rocks typical of orogenic belts. The time relationship to the tectonic phases, however, is very variable: some are synkinematic while others, probably the majority, are late- or even post-kinematic. Trondhjemite is also a common component of ophiolites and may be the only "granitic" rock that is abundant in ocean crust.

Y. K. BENTOR

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Cross-references: *Anatexis*; *Andesite and dacite*; *Batholith*; *Charnockite suite*; *Diorite and tonalite*;

Granite; *Plutonic rocks—classification and nomenclature*; *Rhyolite*.

GRANOPHYRE

A *granophyre* is a felsic, holocrystalline, intrusive or extrusive rock in which quartz and feldspar are abundant and intimately intergrown micrographically in a regular or irregular manner. This type of texture is known as *granophyric* (Fig. 1; Johannsen, 1939).

As with many rock names, the meaning of granophyre has changed considerably since it was first coined by H. Vogelsang in 1872 as a name for porphyritic rocks possessing a holocrystalline granular groundmass with minor interstitial, imperfectly crystallized material. In 1877, H. Rosenbusch used granophyre as a name for porphyritic or nonporphyritic felsic extrusive rocks that have a holocrystalline and granophyric groundmass. It is now used essentially in the Rosenbusch sense, but it has been extended to include intrusive rocks with granophyric texture. Commonly, granophyres are found genetically associated with tholeiitic (basaltic) igneous intrusions. Because of this the term granophyre has been broadened even further to include felsic rocks that do not necessarily have a well-developed granophyric texture but that are closely associated with mafic intrusions. Hence, although it is strictly a descriptive rock name, in some cases it has been used with a genetic connotation.

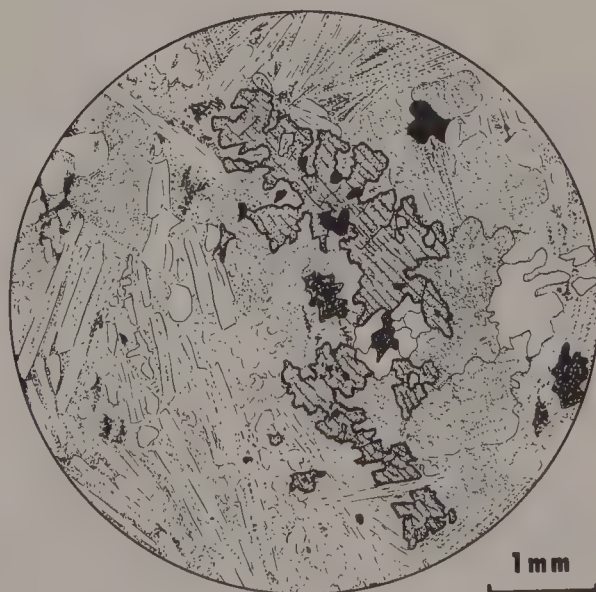


FIGURE 1. Granophyre with a large Fe-rich pyroxene crystal (high relief) set in a granophyric groundmass of colorless quartz (irregular and rod-shaped groups of crystals) and turbid, alkali feldspar; Red Hill dolerite intrusion, Tasmania, Australia.

Granophyres are of diverse origins. Those granophyres associated with tholeiitic (basaltic) intrusions were formed by differentiation in at least some cases (McDougall, 1962). Commonly, these granophyres have a relatively high Fe content and this is characteristic of differentiates in many tholeiitic intrusions. Granophyres may also be produced in basaltic intrusions by assimilation of crustal rocks incorporated at the time of emplacement; some of the more felsic granophyres of the Skaergaard intrusion, Greenland and some granophyres found associated with the Karroo dolerites of S Africa (Walker and Poldervaart, 1949) probably have this origin. Granophyres are particularly abundant in the British Tertiary Volcanic Province, where numerous intrusive stocks, cone sheets, and ring dykes are found in close relationship with intrusions of basaltic magma. In large granitic batholiths, granophyres also occur, commonly as veins and dykes but in larger bodies too. These granophyres are clearly of different genesis from those that are associated with rocks of basaltic composition.

Granophyres found in or adjacent to basaltic bodies of rock generally have pyroxene, or less commonly hornblende, as the ferromagnesian mineral, whereas granophyres in granites are usually characterized by the presence of biotite or muscovite.

Markfieldite is a granophyric rock with phenocrysts of plagioclase together with augite and/or, hornblende in a micrographic groundmass (Markfield, England).

IAN McDOUGALL

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Cross-references: *Assimilation*; *British Tertiary Volcanic Province*; *Hybridization*; *Magmatic differentiation*; *Skaergaard intrusion*; *Textures of igneous rocks*.

GRANULITE

Granulite is a term that arose with separate connotations in different languages and, since its

original definition, has widened considerably in scope and in general use. The most widespread use at present is close to the original German definition of a high-grade metamorphic rock with a granular texture and a strong banding, but possessing no foliaceous minerals, so that a strong schistosity is not developed. This definition of granulite was used by Eskola (Barth et al., 1939) to define the highest grade of regional metamorphic rocks that contain pyroxene in place of normal hydrous ferromagnesian minerals, so that pyroxene granulite has become the most widely known usage. Rocks of this facies (i.e., with a mineral composition in which orthopyroxene is stable rather than common green hornblende) have been recognized throughout the world, being particularly common in areas of Precambrian rocks. The use of the term granulite has now become almost universally synonymous with “granulite facies rock,” and there are now at least three widely recognized subfacies or even separate facies: pyroxene granulite, hornblende granulite, and cordierite granulite (de Waard, 1965), so that the original meaning, which certainly implied rocks with a particular texture, is not always strictly adhered to.

There have also been aberrant usages that bear little or no relation to the original definition. Some French petrologists use granulite for a muscovite granite with accessories such as topaz, apatite, and tourmaline and having a fine granular texture (Termier and Termier, 1956, p. 256). The Geological Survey of Scotland used the term for a metamorphic rock of granulose texture, in which foliation is not marked, and for which the terms schist and gneiss are inappropriate. The term was principally applied to rocks of the Moine assemblage (“Moine granulites”). The principal minerals of these rocks are quartz and feldspars, but no limits to the mineralogy have been proposed. The rocks are of varying metamorphic grade, although never of pyroxene granulite facies, and the name thus created considerable confusion. Unfortunately, the term, which has a long history of usage in stratigraphic formation names (e.g., “Central Highland Granulites”) is still used for such rocks by some authors.

The term granulite originated in the Granulitgebirge of Saxony, where three types can be recognized (Watznauer, 1969): (1) pyroxene-free quartz–feldspar granulite, (2) pyroxene-bearing quartz–feldspar granulite, and (3) plagioclase–pyroxene granulite. Types 2 and 3 are dark-colored, whereas type 1 is the typical light-colored acidic granulite that may have no minerals diagnostic of granulite facies. These acidic rocks are composed largely of feldspar, quartz, and scattered garnets. Minor con-

stituents include biotite, tourmaline, rutile, sillimanite, kyanite, hercynite, apatite, and zircon. The feldspar is microperthitic K-feldspar and antiperthitic oligoclase. The most important character, however, is the texture, which is characterized by thin layers of xenoblastic feldspar and quartz alternating with layers of plate-like quartz grains. The feldspar may also be elongate, but the flattened quartz grains (termed plättung texture) are typical. These acidic rocks are commonly associated with more competent pyroxene-bearing metabasites, ultrabasites, and metasediments together with eclogites. In some of them static recrystallization has often led to a polygonal aggregate of grains whose mutual boundaries commonly meet at 120° triple points.

Studies in the Granulitgebirge, E and W Germany, suggest that there plättung texture and the fine banding is an effect of cataclastic deformation, with recrystallization under anhydrous conditions concomitant with shearing during the tectonic emplacement of a diapiric block. These granulites appear to have formed at a relatively shallow depth, although the presence of kyanite contradicts this, and there is little doubt that the Saxony granulites, although mineralogically of the pyroxene granulite facies, have a different origin from many other granulites. Granulites of the Massif Central and Southern Brittany, France, and possibly those in N Spain are also of this type, and the use of granulite as a textural term to describe such rocks is still prevalent in France and Germany whether or not the mineralogy is diagnostic of the pyroxene granulite facies.

The granulites in the areas of Europe mentioned above have all formed fairly recently, probably in Palaeozoic times, and have been caught up in Hercynian or Alpine orogenies as basement blocks. The other major occurrence of granulite facies rocks is in two major groups of Precambrian rocks, late Archaean to early Proterozoic (about 3,000–2,000 Ma) and mid-Proterozoic to late Proterozoic (about 1,300–900 Ma) (Heier, 1973). These appear to be fundamentally different in origin from Saxony-type granulites, and although many contain plättung texture in the quartz, this appears to be due to extreme flattening deformation most probably not associated with shear structures. The depth of formation is usually assumed to be very great, but many authors also recognize that an anhydrous parent rock can, on metamorphism, give rise to granulite facies assemblages, so that many granulites owe their origin to the metamorphism of igneous bodies or are the end result of a number of metamorphic episodes.

It has been questioned whether rocks of such

disparate origin should be included in the same class of metamorphic rocks (Watznauer, 1969) and several terms have been suggested to replace granulite for those rocks that are not of pyroxene granulite facies, since there is not a widely accepted term for rocks of even grain size and low metamorphic grade.

Goldsmith (1959) suggested the term *granofels* for those metamorphic rocks that are granoblastic without, or with only an indistinct, foliation or lineation. The nomenclature of basic "granulites" would then be by prefixes to the term *granofels*, such as diopside–plagioclase *granofels*. In order to render this less clumsy in the very common cases of multicomponent rocks, Berthelsen (1960) suggested an extension of the type of terminology used in the amphibolite facies: amphibolite is widely used for hornblende–plagioclase rocks of any texture. Berthelsen proposed that the higher-grade metabasites be termed *pyribolite* for pyroxene–hornblende–plagioclase rocks and *pyriclasite* for pyroxene–plagioclase rocks, the end members containing amounts $>\frac{2}{3}$ of the particular component (plagioclase, pyroxene, and hornblende).

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Cross-references: *Amphibolite facies*; *Cataclastic rocks*; *Charnockite*; *Charnockite suite*; *Granulite facies*; *Metamorphic mineral growth*.

GRANULITE FACIES

Granulite facies rocks occur in every continent. They are confined largely to Precam-

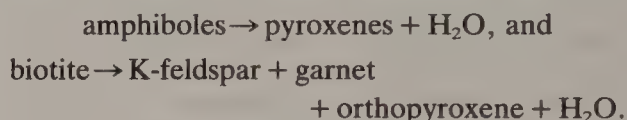
brian shield areas, although they also occur as tectonically uplifted or thrust basement blocks among younger metamorphic rocks (Oliver, 1969; van Breemen et al., 1982). P. Eskola defined the granulite facies in 1939 (see Turner, 1968), the Granulitegebirge of Saxony (E and W Germany) being a type area. The terms granulite and granulitic are much older petrographic-textural terms and this additional use of the term added to the confusion surrounding their varied uses by petrologists of different nationalities. A working group, after considerable discussion, has reached a degree of consensus on petrographic usage. (Mehnert, 1972). A granulite is defined as "a fine to medium-grained metamorphic rock composed essentially of feldspar with or without quartz. Ferromagnesian minerals are predominantly anhydrous. The texture is mainly granoblastic, the structure is gneissose to massive. Some granulites contain lenticular grains or lenticular aggregates of quartz." Granulite is the type rock of the granulite facies, although not all granulite facies rocks should be described as granulites. Simple triple-point mosaic granoblastic textures are characteristic of many granulite facies rocks; in the less competent lithologies, quartz grains are commonly much flattened (plättung texture).

Mineralogy and mineral assemblages

The granulite facies as described by Eskola consists of a range of anhydrous assemblages (Fig. 1). Amongst these minerals, plagioclase is antiperthitic, the alkali feldspar contains up to 50% albite, whilst the pyroxenes may be Al_2O_3 -rich, although the clinopyroxene is jadeite-poor. Rarer but distinctive minerals are scapolite and sapphirine. Olivine, corundum,

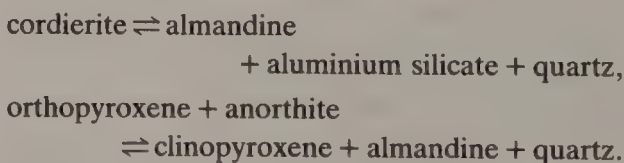
and spinel occur in undersaturated rocks whilst calcite, dolomite, diopside, and forsterite occur in calcareous assemblages. Wollastonite and grossular are found very rarely in these rocks. Rutile, ilmenite, and graphite are characteristic accessory minerals (Droop, 1989).

In many areas, biotite and hornblende, though not muscovite, occur in otherwise anhydrous assemblages. They may result from a retrograde metamorphism, when often a new structural trend is imposed. Alternatively, such assemblages are transitional between the amphibolite and granulite facies. Reactions marking the boundary between these facies are



These are essentially extremely complex dehydration reactions and detailed studies show marked compositional changes in biotite and hornblende, as well as decreasing modal proportions of these minerals, with increasing grade of metamorphism. An orthopyroxene-in isograd within a range of rock types marks the granulite facies boundary. The transitional region where anhydrous and hydrated ferromagnesian minerals coexist has been described as the hornblende granulite subfacies (Turner and Verhoogen, 1960) or facies (de Waard, 1965) or as a transitional facies (Turner 1968). The amphibole- and/or biotite-out isograd marks the boundary of the pyroxene granulite subfacies or facies when the idealized anhydrous assemblages become stable. Hornblende and pyroxene granulite assemblages are often intermingled on a small scale, a testimony to the localized activity of water.

A second set of pressure (P_{TOTAL})-dependent isograd reactions has been suggested by de Waard (1965):



The isograds are thus cordierite-out in pelitic and quartzofeldspathic rocks and clinopyroxene + garnet-in for basic or ultrabasic rocks. Experimental work verifies that the latter reaction will occur at higher pressures. Both reactions are bulk composition-dependent. Wynne-Edwards and Hay (1963) show that the occurrence of garnet or cordierite + garnet in Grenville rocks (Canada) is controlled by FeO/MgO ratios and CaO contents in rocks. Experimental work (Green and Ringwood, 1967) and field observations suggest that the

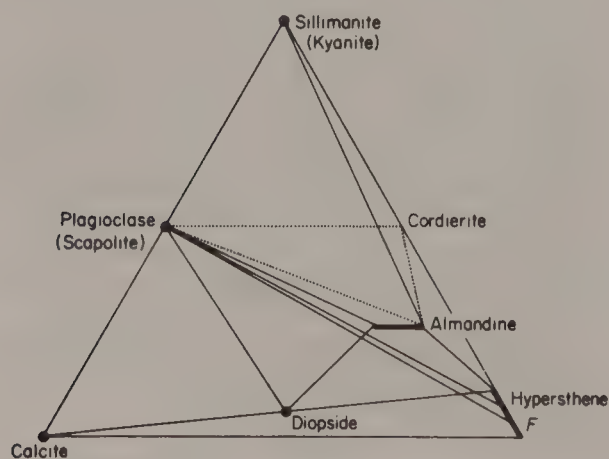


FIGURE 1. Generalized ACF diagram for the granulite facies for assemblages with excess SiO_2 and K_2O . Quartz and K-feldspar are additional phases. (After Turner, 1968.)

garnet + clinopyroxene $\pm$ plagioclase assemblage is formed under less extreme conditions in undersaturated basic rocks. A broad transitional region must be expected for both reactions, within which contrasting assemblages coexist. The granulite facies was divided into six subfacies by de Waard (1965) on a basis of dehydration and load pressure isograds. The complete sequence of subfacies cannot be seen within one area, though individual areas may contain four subfacies.

Igneous rocks

Large anorthositic masses, completely gradational into norites and mangerites, are commonly associated with granulite facies complexes (e.g., the Adirondacks, U.S.A. and the Grenville Province, Canada). These masses are either pre- or syn-orogenic and are frequently deformed. Recrystallization within granulite or sometimes amphibolite facies, plutonic crystallization under granulite facies conditions, and magmatic crystallization under low P_{H_2O} variously account for the rock types seen (see **Charnockite suite**).

Chemical mobility

Many pyroxene granulite facies rocks are strongly depleted in K, Rb, Th and U relative to equivalent amphibolite facies rocks. Any retro-grade metamorphic event (e.g., amphibolite facies metamorphism of granulite facies rocks of the Lewisian complex, NW Scotland) involves increasing K and H_2O although not necessarily related trace elements. The depletion process may result from anatexis and the mobilization of melt or from hydrous fluid migration (Tarney et al., 1972).

Conditions of metamorphism

Various authors (Hietanen, 1967; Turner, 1968) suggest temperatures of $700 \pm 50^\circ\text{C}$ for the initiation of granulite facies conditions. Pressures within the range 5–15 kb seem appropriate. At such temperatures abundant anatectic granites should form if $P_{H_2O} = P$. The relative rarity of such granites suggests that P_{H_2O} is less than P and that the fluids present are CO_2 rich.

In southern Norway Touret and Olsen (1985) put the transition from amphibolite to granulite facies at $700\text{--}800^\circ\text{C}$ and 6–8 kb. These workers are particularly concerned with the study of fluid inclusions and they show elegantly that these granulite facies rocks contain (1) sequentially high density CO_2 , (2) CO_2 of decreasing density with independent pulses of N_2 , (3) CH_4 and hydrocarbons associated with graphite, (4)

CH_4 and H_2O as simultaneous or separate pulses, and (5) fluid inclusions consisting only of H_2O .

Many granulite facies rocks are polymetamorphic.

M. A. LAPPIN

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Cross-references: *Amphibolite facies*; *Anorthosite*; *Charnockite suite*; *Eclogite facies*; *Granulite*; *Metamorphic facies*; *Metamorphic mineral growth*; *Metasomatism*; *Polymetamorphism*.

GREENSCHIST FACIES

Greenschist facies is characterized by an assemblage of actinolite + chlorite + epidote + albite in basic rocks. In pelitic rocks there are some variations and various subfacies have been suggested based on the presence or absence of

biotite. Lower-grade metamorphic facies contain such minerals as zeolite, prehnite, and pumpellyite, which disappear at the greenschist facies boundary (see **Zeolite facies; Prehnite-pumpellyite facies**). At higher pressures glaucophane-lawsonite assemblages grade into greenschist facies assemblages by loss of lawsonite but blue-green amphiboles are stable over a wide transition zone in the “glaucophanitic greenschist” subfacies (see **Glaucophane schist facies**). Presumably eclogites form the high-pressure boundary to greenschists with the formation of omphacite and pyrope-rich garnet, but this progressive transition is not known in nature.

Higher temperatures and pressures lead to either the epidote amphibolite (at higher pressure) or amphibolite (at lower pressure) facies with a change essentially from actinolite to hornblende and from albite to more calcic plagioclase.

Many authors (e.g., Winkler, 1967; Turner, 1981) regard the epidote amphibolite facies as a subfacies of the greenschist facies or as part of a transition zone between the greenschist and amphibolite facies. In this volume it is considered separately (see **Amphibolite facies; Epidote amphibolite facies**).

Basic rocks. It is in basic rocks that the boundaries are defined. The two lower-grade facies are (1) prehnite-pumpellyite and (2) pumpellyite-actinolite schist facies. The boundary of (1) is marked by the breakdown of prehnite and the boundary of (2) by the disappearance of pumpellyite, but the stability relationships are not known in detail. At high grades, actinolite gives way to hornblende, again a reaction for which there is no experimental evidence, although it may well be a transitional change.

The transition from the greenschist to glaucophane schist facies is marked by distinctive transitional subfacies (Banno, 1964—see Turner, 1981, pp. 299–300). Pumpellyite-bearing rocks, with assemblages such as albite-epidote-chlorite-pumpellyite-actinolite, succeed the lawsonite-bearing schists and are succeeded by epidote-glaucophane schists (also containing albite, chlorite, and actinolite).

Pelitic rocks. Two assemblages are widely recognized within the greenschist facies: quartz-albite-muscovite-chlorite (Fig. 1) and quartz-albite-epidote-biotite (Fig. 2). These correspond to the chlorite and biotite zones of Barrow's classic regional metamorphic zonal sequence and may be regarded as identifying subfacies of the greenschist facies.

Within the chlorite zone, McNamara (1965) has suggested that various carbonate-bearing assemblages in pelitic rocks in the Dalradian

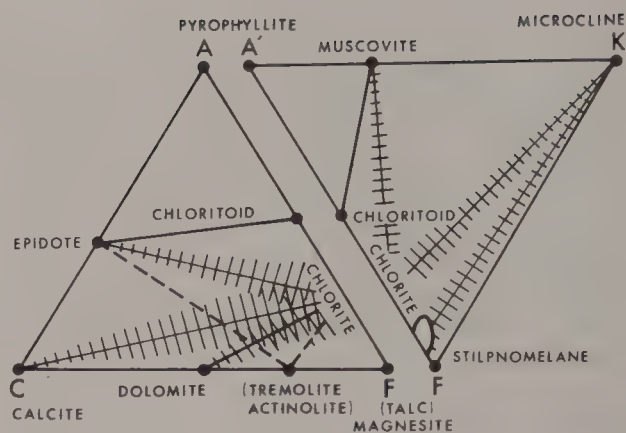


FIGURE 1. ACF and A'KF diagrams of the quartz-albite-muscovite-chlorite subfacies; minerals in parentheses appear only in CO_2 -poor rocks—in such cases the dashed tie lines are valid; variations in composition of chlorite changes tie-lines as indicated by cross-hatching. (After Winkler, 1967.)

Supergroup of the Scottish Caledonides are equivalent to zeolite- and pumpellyite-bearing rocks of the zeolite and prehnite-pumpellyite facies of areas where carbonates are unstable (Fig. 3). Thus, with a high relative chemical potential of CO_2 , dolomite is stable at a lower temperature and calcite at a higher temperature. He suggests that the change from dolomite to calcite also marks the incoming of biotite and the chlorites become more magnesian. These subfacies are disputed by Turner (1981) who suggests that all such subfacies are shown “by experience” to be closely associated together in the field, along with pelitic assemblages of constant character. McNamara's identification of biotite is also disputed. This does not, however, explain the absence of zeolite- and pumpellyite-bearing assemblages from the Scottish Dalradian rocks and their widespread development in many other areas. It seems highly likely that the development of carbonates

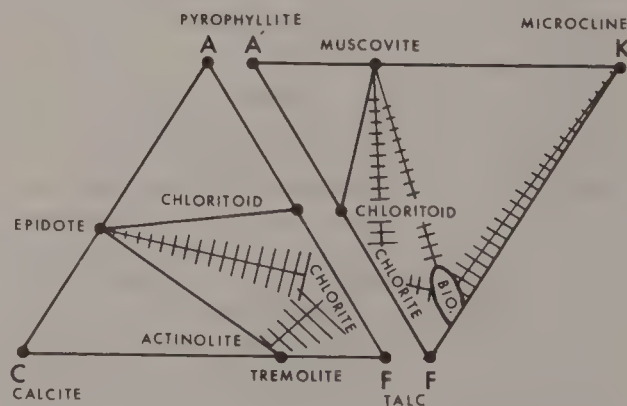


FIGURE 2. ACF and A'KF diagrams of the quartz-albite-epidote-biotite subfacies; variations in composition of chlorite changes tie-lines as indicated by cross-hatching. (After Winkler, 1967.)

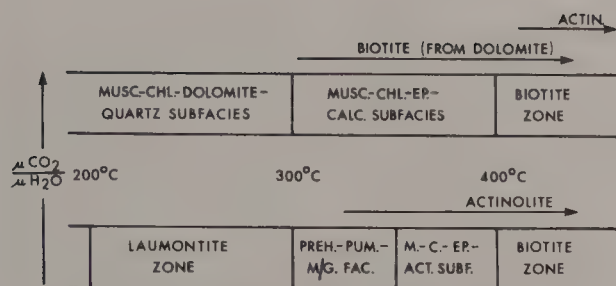


FIGURE 3. Schematic representation of $\mu_{\text{CO}_2}/\mu_{\text{H}_2\text{O}}$ and T conditions for Scottish Dalradian (above) and New Zealand (below) rocks in the lower greenschist facies; other workers suggest that the "biotite (from dolomite)" is an oxidized chlorite: m(usc.), muscovite; c(hl.), chlorite; ep., epidote; calc., calcite; preh., prehnite; pum., pumpellyite; m/g., metagraywacke; act(in.), actinolite. (After McNamara, 1965.)

in pelitic rocks inhibits the formation of the calcium aluminosilicates that otherwise define the lower limits of the greenschist facies in pelitic rocks.

Although the incoming of almandine garnet marks, more or less, the transition to epidote amphibolite facies in pelitic rocks, garnet is quite commonly found in greenschist facies rocks. It is, however, a manganiferous garnet rich in the spessartine molecule, and usually occurs as very small scattered crystals.

Calcareous rocks. Both calcite and dolomite are stable with quartz, and impure calcareous rocks (marls or sandy limestones) have only epidote–zoisite or chlorite commonly developed (in addition to carbonate and quartz) at lower grades. Towards the higher-temperature end of the stability field of this facies, quartz reacts with carbonates and tremolite becomes stable.

Ultramafic rocks. A number of different parageneses are possible and almost any combination of two or three of the following minerals are found—antigorite, talc, magnesite, dolomite, brucite, diopside, tremolite, chlorite, and magnetite (depending on the silica saturation and CO_2 content). The incoming of forsterite is probably fairly close to the biotite isograd in pelitic rocks.

Carbonates (sometimes also including calcite) are very commonly developed but this is naturally due to metasomatic introduction of CO_2 into meta-igneous ultrabasic rocks.

A. E. WRIGHT

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Cross-references: *Amphibolite facies*; *Eclogite facies*; *Epidote amphibolite facies*; *Metamorphic facies*; *Metamorphic grade*; *Prehnite–pumpellyite facies*; *Pumpellyite–actinolite schist facies*; *Regional metamorphism*; *Zeolite facies*.

GREENSTONE

Greenstone is an old field term applied to any compact dark-green, altered, basic to ultrabasic igneous rock. Commonly, chlorite–actinolite (–albite) assemblages represent the products of greenschist facies metamorphism. In other cases, late magmatic fluids have caused autometamorphism—uralitization of pyroxene, serpentinization of olivine, and associated saussuritization of calcic plagioclase.

The term *greenstone belt* is used to describe volcanic-rich assemblages that are common in Precambrian shield regions, particularly in Archaean terrane where greenstone is abundant (Archibald et al., 1978; Condie, 1981). Basalt–andesite–dacite–rhyolite flows and pyroclastic rocks dominate over large areas; ultrabasic rocks (including komatiites) also occur as do metagraywackes, metapelites, and cherty iron-formation. Although now largely consisting of greenschist facies assemblages, the rocks have retained, or largely retained, original textures and structures.

A. M. GOODWIN

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Cross-references: *Archaean volcanism*; *Greenschist facies*; *Ultramafic lavas*.

GREISEN

Greisen is a term used originally by miners of Saxony, E Germany for a variety of granite

containing only Sn ore, mica, and quartz, and little or no feldspar (von Leonhard, 1823). In modern terms it may be defined as a pneumatolytically altered granite composed mainly of quartz, pale silvery-green mica, and topaz. The mica is usually muscovite or a Li-bearing mica such as lepidolite or zinnwaldite. The original feldspar and biotite of the granite are replaced by quartz, Li-bearing mica, tourmaline, and some kaolinite; other minerals that commonly occur in greisen are apatite, fluorite, and ore minerals such as cassiterite, rutile, wolframite, and chalcopyrite. Greisen is a coarse-grained, gray, granular rock, with a silvery glittering appearance by reason of its mica content.

The original definition implies a secondary origin, and most greisens occur as veins up to a few tens of cm wide that pass gradually into the surrounding granite, and that frequently have at their center a fissure filled, or partly filled, with quartz and ore minerals. A few authors, however, prefer a nongenetic definition and use the name for all igneous rocks composed essentially of quartz and white mica; e.g., Johannsen (1932), following Spurr (1906), restricted the term greisen to quartz-white mica rocks produced by the alteration of granite, and used the name *esmeraldite* for those that occur as veins with sharp intrusive contacts and that were produced other than by metasomatic processes.

Greisen (in the restricted sense) is formed by the reaction between granite and vapors or fluids rich in F, B, and Li, the hydrothermal alteration process being called *greisenization* or *greisening* (Hall, 1971). The hydrothermal fluids were probably given off by the cooling granite and the changes they effect follow closely the time of its emplacement. The fluids follow

fissures in the granite and sometimes in the adjacent country rocks and there react with the wall rock to produce greisen (Burt, 1981; Manning and Pichavant, 1983). The fissures may form a well-defined pattern, or they may be so closely intersecting that virtually the whole of the granite is altered to greisen.

Compared with the unaltered granite, most greisens are poorer in Na and richer in F, Li, Rb, Sn, and Zn. The pattern of chemical changes varies from area to area, and individual veins may also show marked variations in the relative proportions of other elements.

Greisen is worked in places as an ore of Sn and W. The old Saxon miners' term for such Sn ore is "Zwitter," and the rock containing the ore is called "Zwittergestein."

A. C. BISHOP

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- Cross-references: *Granite*; *Granites and mineralization*; *Metasomatism*; *Pneumatolysis*; *Volatiles*.

H

HARZBURGITE

Harzburgite was defined by Rosenbusch (1887) as a feldspar-free rock composed of forsteritic olivine with essential amounts of magnesian-rich orthopyroxene, either enstatite or bronzite. Accessory spinel, usually chromite or picotite, may be present, and some harzburgites contain a small amount of chrome diopside.

The name *saxonite* was originally applied to rocks composed of olivine and enstatite, but this name is not now widely used: for a full discussion of the use of the terms *saxonite* and *harzburgite*, see Rosenbusch (1907, p. 465–468) and Johannsen (1938, p. 438–439).

Unserpentinized harzburgite is a dark olive-green, coarse-grained rock weathering red-brown on the surface. The orthopyroxene tends to resist weathering and serpentinization and forms rough projections on weathered surfaces. The rough surface and darker color serve to distinguish harzburgite from dunite in the field. Even slight serpentinization can alter harzburgite to a dark gray or black color, although even completely serpentinized harzburgite may still be recognized by the presence of bronze-colored bastite pseudomorphs after orthopyroxene.

Harzburgites are usually allotriomorphic granular with anhedral, equidimensional grains of olivine and orthopyroxene (3–5 mm) interfering along sutured margins. Euhedral or anhedral grains of picotite (0.1–0.3 mm) may be included

in olivine and orthopyroxene or interstitial to them (Fig. 1). Small interstitial grains of chrome diopside may also be present. The orthopyroxene commonly contains exsolution lamellae of clinopyroxene oriented parallel to (100). In some harzburgites, large oikocrysts of orthopyroxene enclose corroded grains of olivine. Replacement of olivine by orthopyroxene may be marked by irregular clusters of spinel grains which indicate the original outline of the olivine (Fig. 2).

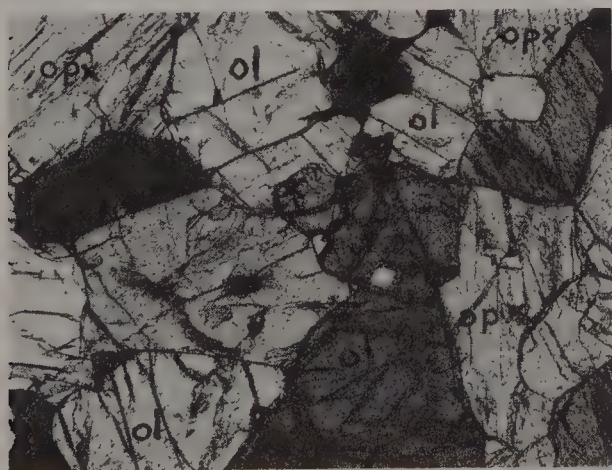


FIGURE 1. Photomicrograph of harzburgite, Great Southern massif, New Caledonia: ol, olivine; opx, enstatite; black, picotite; PPL, $\times 13$.

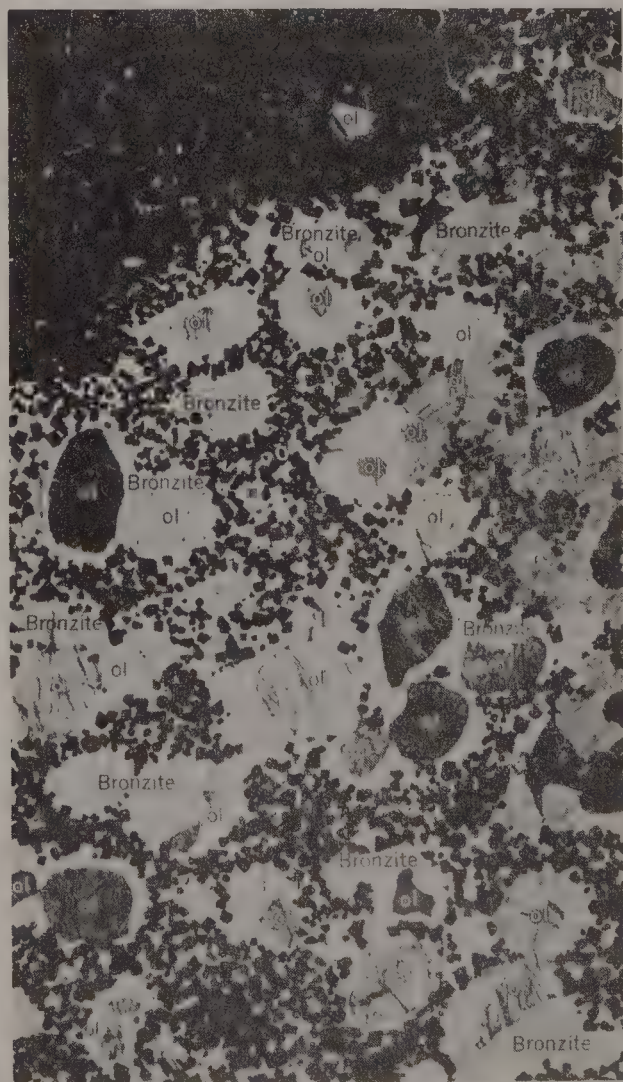


FIGURE 2. Photomicrograph ($\times 6$) of picotite-rich harzburgite, Stillwater intrusion, Montana, U.S.A., showing oikocrysts of orthopyroxene (Bronzite) enclosing and replacing olivine (ol) and an overlying chromitite layer. (From Jackson, 1961.)

Many harzburgites show evidence of deformation in the form of undulose extinction and kink-bands in the olivine, and bent exsolution lamellae in the orthopyroxene.

Chemically, harzburgite is similar to dunite but has slightly higher SiO_2 and lower MgO . Although there is usually less picotite in harzburgite than in dunite, the total Cr_2O_3 content is generally not reduced because of increased amounts in the pyroxene. The chemical analysis of an unserpentinized harzburgite (Red Hill, New Zealand; Challis, 1965) is: SiO_2 43.56, TiO_2 tr., Al_2O_3 0.9, Fe_2O_3 0.68, Cr_2O_3 0.61, FeO 7.25, NiO 0.32, MnO 0.10, MgO 45.25, CaO 0.90, Na_2O tr., K_2O 0.00, $\text{H}_2\text{O} +$ 0.40, CO_2 0.00; trace elements (p.p.m.) are Co 180, Cu 10, V 50.

Harzburgites occur in the mantle, in "alpine-type" ultramafic intrusions and in the basal zones of stratiform layered intrusions (Menzies, 1977; Dawson et al., 1980).

G. A. CHALLIS
W. R. LAUDER

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Cross-references: *Bushveld Complex*; *Intrusion—layered*; *Kimberlite*; *Lherzolite*; *Ophiolite*; *Peridotite*; *Plutonic rocks—classification and nomenclature*; *Serpentinite*; *Ultramafic rocks*; *Upper mantle—experimental petrology*.

HAÜYNITITE

Haüynitite is a plutonic or hypabyssal rock mainly composed of haüyne and pyroxene

(usually titanaugite) and with a small proportion of feldspathoids. A monomineralic effusive rock composed entirely of haüyne is termed a *haüynolith*. Where the composition is similar to a leucitophyre, but with haüyne in place of some of the leucite, or where in a rock like nephelinite the place of nepheline is taken by haüyne and nosean, the term used is *haüynophyre* (*haüynporphyr*). *Gleesite* is a collective name for subvolcanic haüyne rocks.

D. R. BOWES

Cross-references: *Alkaline rocks—undersaturated*; *Feldspathoidal rocks—genesis*; *Leucitite*; *Melilitite*; *Nephelinite*; *Noseanite*; *Sodalitite*.

HORNFELS

Hornfels are hard, fine- to medium-grained and often flinty-looking metamorphic rocks in which color is not diagnostic and depends largely upon the initial composition. They are composed of a mosaic of approximately equidimensional mineral grains that result from the complete, or nearly complete, recrystallization of the constituents present prior to metamorphism; relict structures such as schistosity are indistinct or absent. The mineral components are typically <1 mm in size. Large crystals may be present as porphyroblasts or relict phenocrysts.

Occurrence

Hornfels occur in the inner parts of aureoles surrounding igneous intrusions and as xenoliths. They are products of contact (thermal) metamorphism caused by the heat released from the intrusion. Sanidinite facies hornfels can be formed by the exceptionally high temperatures and low pressures present in basaltic dykes and volcanic necks, particularly in xenoliths. Hornfels are characteristic of the pyroxene hornfels facies of contact (thermal) metamorphism, the temperature conditions of which are reached adjacent to some igneous plutons (particularly basic masses) or in the roof pendants of some magma chambers. Hornfels of the hornblende hornfels facies extend outward from the intrusive body and may grade into the more schistose or foliated spotted schists and slates, rocks less thoroughly recrystallized than the hornfels. They are common in the aureoles of granitic batholiths (see **Aureole**; **Batholith**).

Formation

Recrystallization of sedimentary rocks at low pressure but at a wide range of temperatures

(ca. 300–1000°C) accounts for the formation of most hornfels. Originally, they were considered to be silicified sediments. It was later recognized that hornfels can be produced without large compositional changes and that their composition depends upon the composition of the original rock. A wide variety of compositions are thus possible: this idea, stated in its extreme, can be called the “Kjerulf–Rosenbusch rule” after two early geologists who reached this conclusion. In many cases, nearly isochemical metamorphism does occur and little other than the intrusion heat and the addition or removal of water is involved. However, the more volatile constituents of both the metamorphosed rock and the intrusion can play a major role.

Recrystallization reflects the adjustment of the rock to the new energy requirements of a higher temperature; either a smaller grain size, typical of pelitic rocks, or larger grain size, common in marbles, may result. Mosaic texture and equant grains indicate the general lack of shearing stresses. Higher temperatures reduce kinetic barriers to the formation of a more stable mineral assemblage. Fluids initially in the rock (such as H_2O , CO_2 , HCl) may become chemically more active and cause chemical alteration of the minerals. The intergranular fluid will also act as a solvent and the increased mobilization of the rock constituents, especially through this fluid, will aid the recrystallization process. New mineral phases may or may not form, depending upon the relative stabilities under the temperature–pressure conditions. In general, volatile-bearing minerals such as clays, micas, and carbonates will tend to dissociate at higher temperatures. P_{FLUID} will increase near the intrusion (heat source), giving rise to chemical potential gradients. This causes fluids to move away from the intrusion, unless it is volatile-deficient, depleting the recrystallizing rock in these components.

The volatile fluid released by the cooling magma may invade the surrounding rocks and can have important affects on hornfels formation. This fluid may be chiefly responsible for the transfer of heat and it can contain elements such as C, H, S, Cl, F, and B. The addition of these components to the hornfels causes *contact metasomatism* to accompany contact metamorphism. In some cases, large amounts of the less-volatile elements such as Si, K, and Fe may be added during metasomatism to form *tactites* and *skarns*, and yield a hornfels composition much different than that of the initial rock. Volatile pressures considerably greater than hydrostatic may be developed, especially near the degassing intrusion. At times, P_{FLUID} may exceed P_{LOAD} and cause fracturing.

Compositional controls

The influence of the chemical composition of the initial rock on hornfels mineralogy can be shown by diagrams such as Fig. 1, which indicates the compositional range from shale to siliceous carbonate. These rocks usually contain sufficient SiO_2 for the formation of quartz in addition to other minerals. Differing mineralogies then depend upon other major chemical components (Al_2O_3 , CaO , $MgO-FeO$), which will exhibit marked differences. Ten classes of hornfels can be outlined on the diagram according to the mineral assemblages expected to form from different bulk compositions; class 1 approximates a pure shale and class 10 represents siliceous carbonate. Following the mineralogical phase rule, no more than three minerals in addition to quartz should occur in one of these hornfels; this is shown by the positions of the class numbers. If feldspar is abundant in the initial rock, an alkali mineral may occur in addition. The fact that many natural occurrences closely follow this general scheme shows the dominant control of initial rock compositions, as well as the applicability of the phase rule to natural systems.

Complexities and variations on this scheme can usually be explained as deviations from the more ideal initial compositions, or as the result of volatile interaction with the rock during

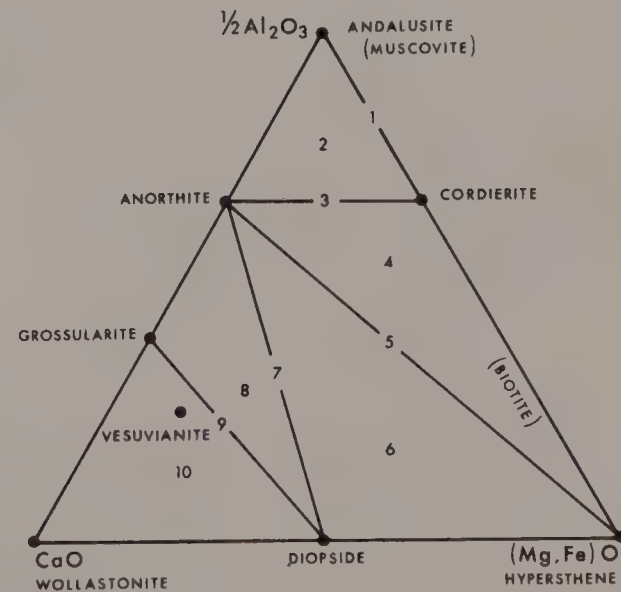


FIGURE 1. The mineral assemblages of Goldschmidt's ten classes of hornfels; numbers on lines or within triangles indicate mineralogies typical of metamorphic rocks formed from sedimentary rock compositions between shale (1) and siliceous carbonate (10); other minerals may also be present if their chemical constituents are abundant; quartz, orthoclase, and minor accessories are common. (After Barth, 1962.)

TABLE 1. Common mineral assemblage of hornfelses^a

Facies	Calcareous				Basic and magnesian
	Pelitic	Quartzo- feldspathic	Calci	Dolomitic	
Sanidine	Cordierite-spinel- mullite	Tridymite-anorthite- glass	Spurrite-calcite Spurrite-larnite	Melilite-monticellite- diopside	
	Cordierite-mullite- glass		Larnite-wollastonite Wollastonite-quartz	Melilite-merwinite- spurrute	
Pyroxene hornfels	Anorthite-corundum- spinel			Monticellite-calcite Monticellite-periclase- calcite	
	Andalusite-cordierite- biotite-orthoclase- quartz	Quartz-orthoclase- plagioclase- biotite	Calcite-wollastonite Calcite-wollastonite- grossular	Calcite-periclase (brucite)	Plagioclase- diopside-
Amphibolite	Sillimanite- biotite-orthoclase- quartz		Wollastonite- grossular-	Calcite-forsterite (Humite)-periclase (brucite)	hypersthene Hypersthene- cordierite
	Andalusite- corundum-spinel		Idocrase-quartz Anorthite-scapolite- grossularite	Calcite-periclase- spinel-forsterite Calcite-forsterite- diopside	Hypersthene- diopside- forsterite- spinel
Albite- epidote amphibolite	Muscovite- biotite-quartz- microcline- plagioclase	Quartz- microcline- plagioclase- muscovite- biotite	Calcite-quartz Calcite-diopside- grossularite	Calcite-diopside- wollastonite	Forsterite- spinel
	Muscovite-biotite- quartz-andalusite		Grossular- plagioclase Calcite-grossular- idocrase	Calcite-brucite forsterite Calcite-forsterite- diopside	Plagioclase- hornblende- diopside Cordierite-
Albite- epidote amphibolite	As for amphibolite facies	As for amphibolite facies	Calcite-quartz Calcite-tremolite- quartz	Calcite-tremolite Hornblende- hedenbergite	anthophyllite (cummingtonite) Biotite- anthophyllite (cummingtonite)
	As for amphibolite facies			Tremolite-dolomite- calcite Forsterite-dolomite Talc-dolomite	Albite- epidote- actinolite

Source: after Williams et al. (1954).

^a Note amphibolite here is used in relation to a facies of thermal metamorphism—most authors would restrict its use to facies of regional metamorphism and use hornblende hornfels and albite-epidote hornfels facies.

metasomatism. Rocks deficient in CaO and/or SiO₂ may display a different hornfels series that is more dependent upon the silica content. Minerals stabilized by F (scapolite, fluorite), B (axinite), P (apatite), or S (sulfides) may also be present. Further complications may come about through the inability of the rock to maintain equilibrium during changing conditions of temperature or volatile pressures. Relict phases (formed earlier during the temperature increase), or late minerals developed during retrograde metamorphism may be found, especially if volatiles could not escape.

Representative mineralogies such as those in Fig. 1 must be considered in terms of the wide range of temperatures and variable volatile pressures under which hornfels can form; hydrous and carbonate minerals are strongly dependent upon both parameters. An example of this is the series of calc-silicate minerals that develop in siliceous carbonates. With increasing temperature and constant P_{CO_2} , a series of Ca-Mg silicates can form, each of which may replace those lower in the series. The rock tends to become continually depleted in CO₂ if avenues for its escape exist. The mineral wollastonite (Fig. 1) is a member of this series and it will form in a hornfels only if the appropriate T - P_{CO_2} conditions are achieved; at other conditions, alternate minerals will be stable. An increase or decrease in P_{CO_2} during metamorphism will inhibit or aid this progressive formation of mineral phases.

Classification

Varieties of hornfels can be named according to certain characteristics of fabric, as in spotted hornfels, or for their characteristic mineralogies, as in biotite hornfels, quartz-feldspar hornfels and forsterite marble. A useful way of classifying hornfels is based upon the compositions of the rocks from which they formed. Four main groups have been outlined by Williams et al. (1954): pelitic hornfels, quartzofeldspathic hornfels, calcareous hornfels, and basic hornfels (Table 1). Of note are the limited number of important mineral phases in any one hornfels, the successive mineralogical changes with increasing metamorphic grade, and the general correspondence with series like the example in Fig. 1.

Other types of hornfels

Astite—mica and andalusite are the main constituents (Cima d'asta, Italian Alps).

Aviolite—mica and cordierite are the main constituents (Aviolo, Italian Alps).

Beerbachite—see *Dolerite*.

Corneite—biotite hornfels with a shale parent.

Cornubianite—fine-grained granulose texture and consisting of micas, quartz and feldspar (Cornubia (Cornwall), England).

Edolite—feldspar and mica are the main constituents (Edelo, Italy).

Halleflinta—fine-grained, compact and granular and formed from an acid igneous rock such as rhyolite.

Hornstone—a general term for a compact, tough, siliceous rock having a splintery or conchoidal fracture such as a hornfels; the term has also been used to describe flint and chert.

Keralite—a quartz-biotite hornfels.

Killas—a term in Cornwall, England, for rocks in the aureoles of granitic intrusions.

Leptynolite—a fissile or schistose variety containing mica, quartz, and feldspar with, or without, andalusite and cordierite.

Proteolite—an early 19th century term for a hornfels-type rock.

Seebenite—feldspar and cordierite are the main constituents.

BERT E. NORDLIE

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Cross-references: *Aureole*; *Batholith*; *Buchite*; *Contact (thermal) metamorphism*; *Contact metasomatism*; *Enclave*; *Facies of thermal metamorphism*; *Hornfelsic textures*; *Metamorphic facies*; *Metasomatism*; *Phase rule*; *Porphyroblast*; *Pyrometamorphism*; *Pyrometasomatism*; *Skarn*; *Volatiles*; *Xenolith*.

HORNFELSIC TEXTURES

A hornfels is a thermally metamorphosed rock; hence, the textures of hornfels are the result of crystallization-recrystallization at elevated temperatures and low to moderate pressures in the absence of directed pressure. Hornfelsic textures are dominated by non-directed features and contain the following textural elements.

Granoblastic (granular)—containing equidimensional, generally xenoblastic, crystals of approximately equal size (Fig. 1).

Polygonal—containing polygonal (five- or six-sided) crystals which tend to have straight or

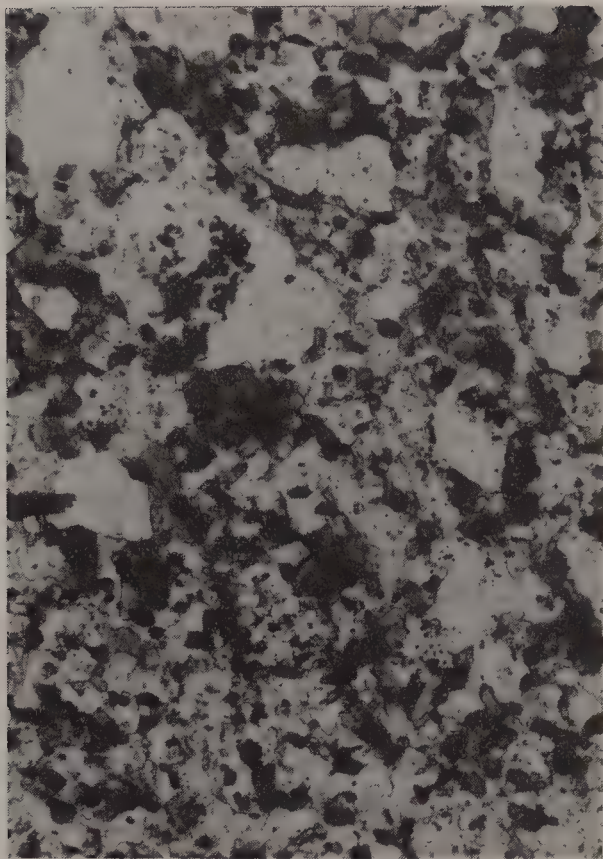


FIGURE 1. Metagraywacke showing original detrital fabric partly obliterated by a granular hornfelsic aggregate of quartz and biotite; PPL, $\times 57$.

gently curved grain-boundaries meeting at triple points.

Decussate—a special case of granoblastic texture in which the crystals tend to be subidioblastic, prismatic or slightly elongate, and have a random orientation.

Idiotopic (“pile of bricks”)—a special case of granoblastic texture in which the majority of crystals are bounded by crystal faces.

Porphyroblastic—containing porphyroblasts; *maculose*, *knotted* or *nodular* rocks contain scattered porphyroblasts (ranging from idioblastic to xenoblastic in shape) in a medium- to fine-grained matrix.

Acicular (needle-like), *fascicular* (bundles of rods), *bow-tie* structures and *spherulitic* aggregates—radially arranged aggregates of elongate crystals.

The dominant *intergranular* textures involve the mutual size, shape, orientation, and distribution of constituent crystals and may be significant within individual crystals (e.g. poikiloblastic and zonal textures, intergrowths, and overgrowths, and reaction rims).

The lack of deformation in contact metamorphism is favorable for the preservation of *relict* (*palimpsest*) textures, which are named by

using “blasto” as a prefix to the name of the texture preserved: e.g., *blastoporphyritic*—it should be noted that true metamorphic textures (e.g., porphyroblastic) involve the use of blastos as a *suffix*. A weak preferred orientation (pseudocleavage) may be formed by mimetic crystallization of mica or amphibole along bedding.

Hornfelsic texture develops in a polycrystalline mineral aggregate by the crystallization of new minerals and the recrystallization of existing stable minerals. Crystallization reduces the total chemical free-energy of the system and recrystallization reduces both the strain-energy within crystals and the grain-boundary energy. Reduction in grain-boundary energy is achieved by an increase in grain size (i.e., decrease in grain-boundary area), the replacement of irregular by regular boundaries, and of high-energy, grain-boundary configurations by lower-energy arrangements. The process is thus analogous to the annealing of metals. The final (“equilibrium,” “stable,” or lowest-energy) texture depends mainly on the relative abundance of the various minerals, on their relative surface energies and surface-energy anisotropies, and on the degree of approach to the lowest-energy configuration (i.e., on operative temperature and time).

Crystallization and recrystallization take place by the initial formation of nuclei that grow through the surrounding grains and finally replace them. The new crystals impinge on each other and grain growth is controlled by the mutual adjustment of grain boundaries. An aggregate of crystals of a single mineral with low structural anisotropy (e.g., quartz) tends to produce a *granoblastic-polygonal* texture with straight, irrational grain boundaries meeting in equal-angle triple points. Minerals with moderate anisotropy (amphibole, wollastonite) produce a decussate texture, and those with high anisotropy (siderite, anhydrite) may produce an idiotopic texture.

Crystallization in an aggregate of minerals with differing surface energy and surface-energy anisotropies (quartz + mica, plagioclase + hornblende, scapolite + diopside), such that one tends to be xenoblastic and the other idioblastic to subidioblastic, involves a compromise between several competing processes. The texture is controlled by the most abundant mineral (i.e., that with the most abundant grain boundaries), not necessarily that with the greatest surface energy. A quartz–mica hornfels can be regarded as either a polygonal aggregate of quartz with mica flakes placed along randomly oriented quartz–quartz boundaries or as a randomly oriented aggregate of mica with interstitial quartz whose boundaries have been pinned by

the mica. Angles at triple points between crystals of the same mineral are equal if the surface-energy anisotropy is low (quartz), but are unequal if the anisotropy is high (amphibole), or if different minerals meet at the triple point.

ALAN SPRY

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Cross-references: *Contact (thermal) metamorphism; Facies of thermal metamorphism; Hornfels; Metamorphic mineral growth; Nucleation of metamorphic minerals; Porphyroblast.*

HOT GROUND

Hot ground is a surface indication of thermal activity closely related to fumaroles. The phenomenon is very common in high-temperature areas and occurs where the steam flow to the surface is so small and dispersed that individual fumaroles or mud-craters cannot be formed. The upflowing steam condenses partially or wholly below the surface, resulting in a hot ground surface. Temperatures ranging from slightly above normal up to 100°C are observed at the surface. The heat flow escaping through the surface of hot ground is, in many cases, of the order of 10^2 to 10^3 (J/m²)/sec, indicating that the steam supply may be of the order of 0.1 to 1.0 (g/m²)/sec. In the case of very active hot ground, the steam condenses within a few centimeters of the surface. Chemical reactions take place in the condensation zone, where the temperature is close to 100°C. H₂S is partially oxidized into elemental sulfur, which is deposited in this zone. A layer of elemental sulfur is formed resulting in a slight heaving of the ground. These small sulfur-heaves are characteristic of the active hot ground. As the sulfur is also being continuously removed by further oxidation and evaporation, the equilibrium thickness of the sulfuric layer generally does not exceed 10–20 cm. In large, high-temperature thermal areas the integrated area of hot ground may amount to many tens of

hectares. Hence, from the energetic point of view, the hot ground is an important type of fumarolic activity. A considerable part of the heat flow dissipated by high-temperature thermal areas escapes through the surface of the hot ground.

G. BODVARSSON

Cross-references: *Fumarole; Geyser; Hot spring; Solfatara; Thermal activity; Volcanoes and volcanism.*

HOT SPRING

Hot spring is a term that can refer to any type of natural outlet for thermal water, steam, or thermal gases on the surface of the Earth. More exactly, the term refers to a natural spring (thermal spring) issuing only, or predominantly, water. The temperature of the water may range from slightly above the normal average for surface water in the region (warm spring) up to the boiling point for water at the altitude of the spring.

Hot water springs are known in practically all thermal areas around the globe. The greatest concentration of them is present in active regions such as Iceland, Yellowstone Park (Wyoming, U.S.A.), and New Zealand. They are commonly associated with steam vents from which superheated steam is rapidly and violently expelled and with mudpots (containing boiling mud) that are usually sulfurous, and often multicolored (paintpots).

The most voluminous hot water spring flows are up to 1000 kg/sec. The average hot spring is much less prolific, with the average of 600 springs in Iceland being about 3 kg/sec.

The water issued by hot springs contains between 10^2 and 10^5 p.p.m. of dissolved solids, mainly SiO₂, NaCl, Na₂SO₄ and other common inorganic compounds. The water may also carry dissolved gases, mainly N₂, CO₂, H₂S and a number of trace gases. The SiO₂ is commonly deposited as an encrustation of siliceous sinter.

There is now a consensus that most of the water issued by hot springs is recycled meteoric water. Juvenile or magmatic water is most likely to be found in postvolcanic fumaroles or solfataras (see Berger and Bethke, 1985).

G. BODVARSSON

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Cross-references: *Fumarole*; *Geyser*; *Hot ground*; *Solfatara*; *Thermal activity*; *Volcanoes and volcanism*.

HYBRIDIZATION

Hybridization is the process of mixing two primary magmas to produce a secondary magma whose products of crystallization are termed hybrid rocks. Early ideas of R. Bunsen, who considered that there were two primary magmas (acidic and basic), were developed by Durocher (1857). In particular, intermediate rocks—including syenites, trachytes, and porphyries—were interpreted as having properties of two primary magmas and referred to as “roches hybrides.”

Added to the concept of the mixing of two magmas were those of impregnation of solid rock by magma and consequent two-way reaction and the partial or complete incorporation of xenoliths into magma (Harker, 1900; see Wager et al., 1965). These are now generally referred to as assimilation (or contamination) and not hybridization. However, some authors have used the term hybridization where the incorporated material was igneous rock and assimilation where it was of sedimentary derivation (cf., Joplin, 1959, and references therein).

The presence of disequilibrium xenocryst assemblages has been used as a criterion for the recognition of the products of hybridization, e.g., (1) the *marscoite* of Skye, interpreted by Harker (1904, see Wager et al., 1965) as a hybrid due to the mixing of an acidic magma containing phenocrysts of quartz and K-feldspar with a basic magma containing phenocrysts of andesine (Wager et al., 1965), and (2) products of the Torfajökull volcano, Iceland, in which Ca-rich plagioclase and Mg-rich olivine coexist with alkali-rich plagioclase and K-feldspar (McGarvie, 1984). Another criterion used is that of isotope systematics; for example, using Pb/Pb, $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$, Dickin (1981) showed that granites in Skye were formed by mixing of acidic magma that was produced by crystallization differentiation of mantle-derived basic magma with partial melts derived from

upper crustal rocks—the latter contributed up to 33% to the Redhills epigranite while *marscoite* requires an 18% contribution from an upper crustal source.

The term *marscoite* was intended for local use only, as was *glamaigite*, as an intrusive rock composed of dark patches of *marscoite* in a lighter-colored groundmass. The term *ringite* for a rock formed by the mixing of silicate and carbonatite magmas appears to have no such local restrictions.

The chilling of a later (usually acidic) magma against an earlier (usually) basic one is a common phenomenon in composite sills and dykes. In many cases, however, the magmas intermingle and hybrid rocks result (see Walker and Skelhorn, 1966; McGarvie, 1984).

C. D. GRIBBLE

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Cross-references: *Assimilation*; *Dyke (dike)*; *Liquid immiscibility*; *Magma*; *Sill*.

HYDROTHERMAL ALTERATION OF SILICATE ROCKS—GENERAL PRINCIPLES—See Vol. IVA

HYDROTHERMAL PROCESSES

Processes operating in nature are called *hydrothermal* if they involve hot water. Their

lower temperature limit is rather ill-defined. Some hydrothermal ore deposits were certainly formed below 100°C. On the other hand, the interaction of sea water with sediments, even at temperatures as high as 45°C in evaporitic basins, is not considered a hydrothermal process. A rough lower temperature boundary for hydrothermal processes is drawn here somewhere between 50° and 100°C.

Water is almost ubiquitous in the Earth's crust, and temperatures above 50°C prevail in most continental areas and below the ocean floor at depths greater than 2–3 km. Hydrothermal processes therefore accompany (1) the transformation of sediments into sedimentary rocks (diagenesis) and of sedimentary rocks into metamorphic rocks, (2) the formation and crystallization of magmas, and (3) the interaction of meteoric, connate, and ocean water with igneous rocks during their rise, solidification, and cooling in the upper portions of the Earth's crust. The quantity of water involved is frequently of the order of km³, although even these rates of transfer are small compared to the enormous volume of water ($\sim 5 \times 10^5$ km³/yr) that is involved in the hydrologic cycle.

Hydrothermal processes during diagenesis and metamorphism

Most sediments are marine. During their burial below the sediment–sea-water interface, reactions between trapped sea water and solid particles occur, particularly in sediments rich in organic matter. In these sediments SO₄²⁻ in trapped seawater is reduced to S²⁻, which is frequently precipitated as a constituent of one or more of the iron sulfides, generally below 50°C (Berner, 1980). The first major reaction above 50°C tends to be the loss of interlayer water from clays of the montmorillonite group near 110°C, and the subsequent breakdown of these clays to illite and chlorite. However, the attainment of equilibrium between minerals in sediments remains incomplete until temperatures considerably in excess of 110°C are reached.

During burial, the porosity of sediments decreases continuously. The salinity of water expelled initially tends to be similar to that of the overlying water column. Toward greater depth, interstitial waters are frequently more saline, and the relative proportions of the cations and anions change in response to a variety of diagenetic reactions. Among these, dehydration, decarbonation, and redox reactions are particularly important. The dewatering of sediments in marine basins is of great importance for the movement of petroleum, and is almost certainly a critical factor in the

development of Mississippi Valley-type Pb and Zn ore deposits.

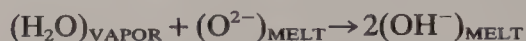
With increasing temperature, mineral assemblages in metamorphic terranes become progressively dehydrated, carbonate minerals lose CO₂ and are gradually converted to oxides and/or silicates, and pyrite reacts with iron oxides and silicates in the presence of elemental carbon to form pyrrhotite and CO₂. The sequence of these reactions is complex, particularly during the progressive metamorphism of shales. The composition of fluids in equilibrium with rocks undergoing metamorphism both responds to changes in mineralogy and can influence the development of the mineralogy where the ratio of fluid to solid phases is sufficiently large (Fyfe et al., 1978; Ferry, 1983). In most metamorphic fluids Na⁺ and Ca²⁺ are the major cations; Mg²⁺ is normally present in very minor amounts; the concentration of K⁺ tends to increase relative to that of Na⁺ with increasing temperature. Cl⁻ is almost invariably the dominant anion; HCO₃⁻, SO₄²⁻, and HS⁻ are normally present in subsidiary quantities. The salinity of metamorphic fluids tends to lie between 1 and 10 wt.% NaCl, but highly saline brines have been found in some fluid inclusions (Rich, 1979). These brines contained up to 40 wt.% NaCl at the time of entrapment, and may owe their high salinity to the presence of halite in sedimentary sequences prior to metamorphism.

The quantity of volatiles released from terranes undergoing metamorphism can be very large. On average, shales contain ca. 5 wt.% water, while their high-grade metamorphic equivalents contain <1 wt.% water. Aqueous solutions generated during metamorphism leave their source regions with a variety of dissolved salts, including those of base and precious metals. Gold deposits in greenstone belts and the gold ores of the Homestake Mine in South Dakota, U.S.A., probably owe their origin to the precipitation of Au from metamorphic fluids. In fluids generated during low-temperature metamorphism the CO₂/H₂O ratio is generally <0.1. At higher temperatures, much larger CO₂/H₂O ratios can prevail, and the fluid phase in at least some granulites appears to consist largely of CO₂ at pressures of several kilobars.

Hydrothermal processes associated with magmatism

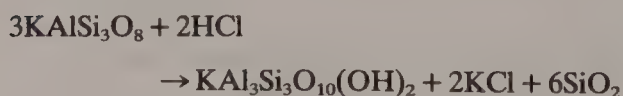
Melting of silicate rocks takes place in the lower parts of the Earth's crust and in the upper parts of the mantle. Water is quite soluble in silicate melts. At water concentrations <ca.

2 wt.% H₂O dissolves largely via the reaction

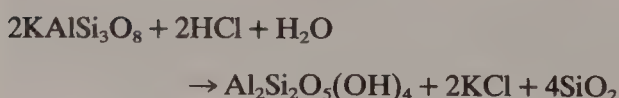


At higher concentrations, water is largely present in molecular form (Stolper, 1982). Approximately 4 wt.% of water dissolves in granitic melts in equilibrium with a vapor phase containing 1 kb of water vapor (Burnham, 1979). At $P_{\text{H}_2\text{O}}$ equal to the lithostatic pressure at the base of the crust, granitic melts can contain in excess of 10 wt.% water. Water therefore tends to move into granitic melts, largely at the expense of hydrous minerals in the area of melting. There is still a considerable amount of dispute regarding the actual concentration of water dissolved in granitic magmas, but the available experimental data on the stability of micas and amphiboles suggest that most granitic magmas contain several wt.% water (Holland, 1972). Most of this must be lost during crystallization, since the quantity of water trapped in hydrous minerals in granitic rocks is usually ≤ 1 wt.%. Granitic melts frequently contain sufficient Cl^- that aqueous fluids released during crystallization are often highly saline. They also tend to be saturated with respect to pyrite or pyrrhotite, and contain sufficient H₂S and SO₂ to lead to the precipitation of a large variety of sulfides during their rise to the surface. Boiling of hydrothermal solutions seems to occur frequently around shallow intrusions. The salinity of fluid inclusions from numerous porphyry copper deposits is highly variable, and extremely high salinities (of the order of 40 wt.% NaCl) are common (Roedder, 1984). During their rise to the surface, such hydrothermal solutions normally produce extensive zones of alteration in the adjacent country rocks. At the highest temperatures, K⁺ metasomatism prevails; at lower temperatures, hydration and neutralization of HCl predominate. Replacement of albite by K-feldspar lowers the K⁺/Na⁺ ratio of the solutions and replacement of plagioclase containing sizeable quantities of anorthite depletes the SiO₂ content of the system (Orville, 1963, 1972). Where SiO₂ cannot be supplied sufficiently rapidly, corundum may form locally.

At temperatures below ca. 500°C, feldspars are usually altered to sericite or kaolinite by reactions such as

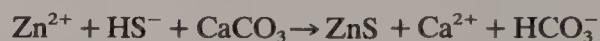


and



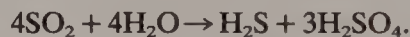
(Shade, 1974). Both increase the pH of hydrothermal solutions. The combination of decreasing temperature and increasing pH increases the degree of saturation with respect to base metal sulfides (Barnes, 1979) and is probably responsible for the deposition of sulfide minerals in many of the major hydrothermal ore deposits surrounding granitic intrusions.

Similar reactions in other wall rocks have much the same effect on the pH of rising hydrothermal solutions. In limestones and dolomites, the replacement of carbonates by sulfide ore probably involves reactions of the type



as well as the direct dissolution of CaCO₃ and CaMg(CO₃)₂.

In relatively oxidized granitic magmas, SO₂ is an important constituent of hydrothermal fluids evolved above 600°C. Between 600° and 400°C, SO₂ reacts with H₂ to form H₂S and H₂O. In the absence of H₂, SO₂ reacts with H₂O to form H₂S and H₂SO₄:



The sulfuric acid generated in this reaction is neutralized by reactions similar to those described above; this process probably accounts for the abundance of anhydrite in many high-temperature sulfide ores associated with granitic intrusions.

By no means all of the water in hydrothermal ore deposits associated in time and space with granitic intrusions was derived from the magmas themselves. Sheppard et al. (1971) have shown that the isotopic composition of hydrogen in secondary clay minerals in the vicinity of many Tertiary hydrothermal ore deposits reflects that of hydrogen in present-day local rainwater, and that the isotopic composition of oxygen in these minerals appears to have been influenced by the value of $\delta^{18}\text{O}$ in the same waters. This is not surprising. Horizontal and vertical thermal gradients around intrusions frequently generate hydrothermal systems that are dominated by meteoric water. The thermal systems at Wairakei in New Zealand and Yellowstone Park, Wyoming, U.S.A. have been studied particularly intensively (Ellis and Mahon, 1977). Essentially all of the water in these systems is meteoric, and the dissolved solids are leached from the volcanic rocks through which the hydrothermal fluids circulate (Ellis, 1973).

The relative proportions of magmatic and meteoric water participating in the formation of hydrothermal ore deposits associated with

TABLE 1. Composition^a of natural hot waters

	1	2	3	4	5	6	7	8	9	10	11	12
T°C	100	100	76	220	216	221	99	250	268	340	195	350
Depth (m)	0	0	0	880	650	300	0	830	775	1,660	447	0
pH (20°C)	1	1.2	2.1	7.8	9.6	5.2 ^b	8.0	8.3	8.4	4.7	9.0	4.0
Li	—	—	2.5	10.2	0.3	—	10.8	13.2	11.2	215	4.5	7.0
Na	7,670	114	395	1,525	212	10,440	1,070	1,250	1,075	50,400	1,280	10,800
K	1,000	65	47	176	27	1,382	102	210	227	17,500	135	1,000
Rb	—	—	—	—	0.04	—	1.1	2.9	2.9	135	0.0	0.2
Cs	—	—	—	—	<0.02	—	2.7	2.5	1.6	14	0.3	—
Mg	7,310	83	10	—	0.0	8	0.4	0.04	0.1	54	0.11	~0
Ca	2,560	210	15	50	1.5	1,812	26	12	3	28,000	4.5	860
Fe	11,480	105	—	—	0.1	—	—	0.05	0.2	2,290	0.13	100
Al	2,030	158	—	—	—	—	—	0.05	—	—	—	—
F	870	100	—	6.6	1.9	0.2	6.6	8.4	8.4	15	23.7	—
Cl	61,840	3,240	511	2,675	197	20,745	1,770	2,210	1,750	155,000	117	19,000
Br	40	8.5	—	—	0.45	—	4	5.5	6.3	120	1.2	—
I	6	—	—	0.2	0	—	0.7	0.3	0.6	18	—	—
SO ₄ ²⁻	—	—	—	—	—	—	—	—	—	—	—	—
As	2,340	1,330	1,215	120	61	72	26	28	2	5	770	~0
CO ₂ (total) ^c	6.3	3	—	—	—	—	2.5	4.5	5.7	12	39	—
SiO ₂ (total) ^d	—	—	59	55	55	2,650	55	17	88	100	1,340	—
B (total) ^d	177	324	330	430	480	374	—	670	830	400	325	1,320
	6.4	23	18	102	0.6	11.6	21.9	288	53	390	26.2	—

^a Concentrations measured in mg/kg, all except 12 at atmospheric pressure from discharge.

^b pH in this sample measured at 221°C.

^c CO₂ (total) = dissolved CO₂ + CO₃²⁻ + HCO₃⁻.

^d For SiO₂ and B, the figures refer to total dissolved silicate and borate species.

1. 1933 crater, White Island, New Zealand.
2. Tamagawa spring, Japan.
3. Spring 9, Rotokaua, New Zealand.
4. Drillhole 2, Rotokaua, New Zealand.
5. Drillhole G-3, Hveragerdi, Iceland.
6. Reykjanes drillhole 2, Iceland.
7. Champagne Pool, Wairakei, New Zealand.
8. Drillhole 24, Wairakei, New Zealand.
9. Drillhole 8, Broadlands, New Zealand.
10. Drillhole I.I.D.-1, Salton Sea, U.S.A.
11. Drillhole KD1a, Kizildere, Turkey.
12. E Pacific Rise at 21°N.

1-5, 7-11 from Ellis (1973); 6 from Björnsson et al. (1972); 12 from van Damm et al. (1983, in Rona et al., 1983).

granitic rocks is still in dispute (Taylor, 1979). It has been shown that meteoric water probably did not penetrate into the hydrothermal system at Casapalca, Peru, until near the close of the period of ore deposition. On the other hand, Ohmoto and Rye (1970) found no trace of magmatic water in the isotopic composition of water in the solutions that formed the base-metal ores at the Bluebell Mine in British Columbia, Canada. The hydrology of magmatic-hydrothermal systems is still somewhat obscure, but it is likely that magmatic water dominates the early evolution of such systems within, and in the vicinity of rather "wet" magmas, and that groundwaters become progressively more important during their later history.

Groundwaters involved in hydrothermal systems are not necessarily derived from local meteoric water. In at least one of the hydrothermal systems of Iceland, seawater is the circulating fluid (Kristmannsdottir, 1983, in Rona et al., 1983). The heavy-metal content of circulated seawater in Iceland is apparently very small, but it is likely that the base-metal deposits of the Kuroko type in Japan and elsewhere owe their origin to seawater circulating through volcanic rocks (Matsukuma and Horikoshi, 1970; Ohmoto and Skinner, 1983). Metals are apparently leached from the country rocks, SO_4^{2-} is reduced to S^{2-} , and base metal sulfides are deposited. The isotopic composition of sulfur in these deposits is most easily interpreted in terms of a marine parentage modified by isotopic fractionation accompanying partial reduction of SO_4^{2-} to S^{2-} .

The world's most extensive hydrothermal systems are entirely submarine. The widespread magmatic activity along mid-ocean ridges generates strong vertical and horizontal thermal gradients in fractured and faulted crust. Seawater enters the flanks of mid-ocean ridges, becomes heated in its downward passage, and emerges at the sea floor close to ridge axes (Ballard and Francheteau, 1983; Sleep 1983; both in Rona et al., 1983) at temperatures close to 350°C (Von Damm et al., 1983, in Rona et al., 1983). Where mixing with cold seawater takes place below the rock-water interface the exit temperatures may be only slightly greater than that of ambient seawater. The warm springs near the Galapagos spreading center are good examples of such mixtures (Edmond et al., 1979).

Seawater that has reacted with basaltic rocks at temperatures of 200°–400°C is typically depleted in Mg^{2+} and SO_4^{2-} and enriched in Ca^{2+} , K^+ , Li^+ , Rb^+ , Fe^{2+} , Mn^{2+} , Cu^+ and Zn^{2+} (Table 1). Na^+ is removed in some systems and is slightly enhanced in others. The same is true

for Cl^- . These differences are perhaps related to the water/rock ratio along the path of seawater through mid-ocean ridge hydrothermal systems (Mottl, 1983). The quantity of seawater cycled annually through basalts is sufficiently large to affect the cation balance and the isotopic composition of seawater (Taylor 1983; Thompson 1983—both in Rona et al., 1983; Holland, 1984). It now looks as if this process is responsible for a good deal of the Mg^{2+} output from the oceans; it may also be an important loss mechanism for marine SO_4^{2-} . It is probably responsible for the excess Li in sedimentary as compared to igneous rocks, and probably buffers the isotopic composition of oxygen in seawater (Holland, 1984).

The role of hydrothermal processes as a source of power and of chemicals has increased considerably in the last three decades. Geothermal power has become progressively more attractive as a source of inexpensive and relatively clean energy, and is a potential major energy source in areas of current and recent volcanism. Nevertheless, it seems unlikely that hydrothermal systems will become a dominant source of power on a worldwide basis (Häfele, 1981).

HEINRICH D. HOLLAND

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Cross-references: *Granites and mineralization; Neutrality-pH and temperature variation; Ocean-floor metamorphism; Pegmatite; Volatiles; Zeolite facies.*

HYDROTHERMAL SOLUTIONS—See Vol. IVA

HYPABYSSAL ROCKS

H. Rosenbusch divided igneous rocks into three groups—*tienfengesteine* (rocks crystallized deep down in the Earth's crust), *erussgesteine* (rocks crystallized from molten magma at the surface) and *ganggesteine* (rocks occurring as

dykes). Later these groups were referred to as plutonic, volcanic, and hypabyssal, respectively. Included in the hypabyssal group are not only dyke rocks, but those occurring in sills, necks, plugs, laccoliths, veins, and apophyses, i.e., those occurring in minor intrusions. They are common in volcanic regions, but apophyses and certain dykes and veins are typically associated with plutonic bodies. The textures of rocks in these two associations differ and are characteristic of the particular environment.

Plutonic rocks are normally coarse-grained, holocrystalline, and granular, whereas volcanic rocks are commonly fine-grained or aphanitic and porphyritic, and, though many are holocrystalline, they may also be holohyaline or hypohyaline and fluidal fabric is often present. The ideas of depth of emplacement and of texture are closely linked and this concept has led to some confusion with regard to the hypabyssal rocks because they may occur in small intrusions at great depth or near the surface where their textures are very similar to those of volcanic rocks. The grain-size of hypabyssal rocks may be fine (small dykes near the surface), is commonly medium and sometimes coarse (very large sills or sheets). Many hypabyssal rocks are porphyritic and some have very typical textures such as the ophitic or subophitic texture of dolerites and the panidiomorphic granular texture of lamprophyres. Dykes, veins, and apophyses associated with plutonic rocks are commonly even-grained and granular and a few are porphyritic. Common rock types associated with volcanic regions are dolerite and lamprophyre, and with plutonic regions, aplite, pegmatite, porphyry, and lamprophyre.

Some petrologists at present regard the term hypabyssal as obsolete and prefer to refer to rocks of minor intrusions.

G. A. JOPLIN

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Cross-references: *Aplite; Cone sheet; Dolerite; Dyke (dike); Intrusion; Laccolith; Lamprophyre; Lopolith; Pegmatite; Phacolith; Plutonic rocks; Porphyry; Ring dyke; Sill; Textures of igneous rocks; Volcanic neck; Volcanic rocks.*

* Major source of references.

IGNEOUS INTRUSIONS— STRUCTURAL BEHAVIOR

Depth of emplacement

There has been a general tendency to equate variations in the form and internal and external structural relations of igneous intrusions with depth of emplacement. Thus, Buddington (1959) suggested a threefold division of granitic intrusions.

1. "Plutons of Epizone"—thought to have been emplaced at depths down to 6–9 km; they tend to be structureless and to have discordant relations to the country rocks that show few if any effects of metamorphism.
2. "Plutons of Mesozone"—thought to have been emplaced at depths of 6–12 km; they are characterized by contact relations that may be in part concordant and in part discordant, by frequently containing well-developed internal structures, and by intruding country rocks that are often strongly deformed near the intrusion and frequently consist of relatively low-grade (epidote amphibolite facies) regionally metamorphosed rocks.
3. "Plutons of Catazone"—thought to have been emplaced at depths below 12–15 km; they are characterized by concordant, and often migmatitic contacts, by the widespread development of internal structures which are generally conformable to the country rock structures, and by country rocks displaying a high grade (amphibolite facies) of regional metamorphism.

However, there is no sharp division between any of these groups and, in addition, intrusions that were apparently emplaced at approximately the same depth at a particular locality may collectively show the characters of all of Buddington's groups, e.g., the granitic intrusions of Donegal (Pitcher and Berger, 1972).

In view of these ambiguities, it would seem that there are no wholly reliable criteria (the relation between the grain-size of an intrusive igneous rock and the depth of emplacement may also be ambiguous) that can invariably be used to assess the level of emplacement of an

intrusion. In consequence, detailed study of an intrusion often provides data that are of more value in enabling issues such as the general relations between magma type and intrusive activity, or the mechanism of emplacement of the magma, to be assessed.

Magma type and intrusive activity

The structural behavior of magma when it is intruded into the crust is dependent on such factors as the contrast in physical properties (particularly density and viscosity) between the intrusive and intruded material, the presence of structural weaknesses in the country rock and the action of regional stresses. Although a general review of intrusive rocks suggests that the interrelationships between the various physical and structural parameters is normally complex, it is possible to recognize certain broad relationships between the mode of intrusion and the physical properties of magmas of different bulk composition.

Thus, it is apparent that basic magma tends to form subhorizontal discoid, sheet-like or funnel-shaped intrusions in many geologic environments. Although this could be explained as being due to the magma exploiting subhorizontal planes of weakness in the country rocks in some instances (particularly at shallow depths in recently deposited sediments), in others, particularly where larger deep-seated bodies are concerned, it is obvious that the intrusions are not being emplaced simply under external structural control. As the basic magma is often as dense as or more dense (ca. $2.7\text{--}2.8\text{ g/cm}^3$) than the crustal rocks into which it is being emplaced ($2.2\text{--}2.9\text{ g/cm}^3$), it seems probable that the subhorizontal character of these intrusions reflects the decrease in "buoyancy" of the magma as it ascends from regions where it has a lower density than the surrounding rocks (possibly as high as 3.6 g/cm^3 if the magma has ascended from the mantle) into the crust, where the reverse relationship holds.

A rather different relationship can be recognized when acidic intrusions are examined. Consideration of the density relations would suggest that such intrusions (density ca. 2.4 g/cm^3) should be more widespread than basic intrusions at shallow depths. However, it

has long been established (e.g., Daly, 1933) that basic rocks are more abundant than acidic rocks in high-level intrusions (and as extrusions) and it is obvious that factors other than density must determine the level to which acidic magma can normally ascend in the crust. An examination of the solidus and liquidus curves of anhydrous and hydrous acidic magma (Harris et al., in Newall and Rast, 1970) provides the probable explanation, namely, that hydrous acidic magma in many instances will be incapable of reaching a low-pressure environment in a fluid state, ascent having led to the boiling-off of water from the magma with a consequent rise in viscosity and in the temperature at which consolidation is complete. Support for this inference is provided by field data indicating that many acidic intrusions were formed as the result of the emplacement of highly viscous and largely crystalline magma. An added implication is that only relatively hot, anhydrous acidic magmas are likely to ascend to high levels in the crust.

Mechanism of emplacement

Consideration of this problem generally resolves itself into a discussion of the means by which preexisting rocks have been displaced in order to accommodate the magma. Two main types of emplacement (displacement) mechanism have been recognized, forcible (forceful) and passive (permitted).

Forcible (forceful) intrusions. In these intrusions the magma has pushed the country rocks aside during emplacement. The most readily recognizable examples of this type of intrusion consist of granitic rocks which would be grouped by Buddington in his "Plutons of Mesozoic." These intrusions are surrounded by zones of deformation in which the country rock structures are deflected into parallelism with the boundaries of the intrusion (Fig. 1) and in which there is often widespread evidence of strong shearing movements. Internally, most intrusions of this type contain a planar structure defined by oriented crystals (predominantly of the minerals of early formation, e.g., plagioclase, hornblende, biotite), xenoliths and basic inclusions (Fig. 2). This structure is normally oriented parallel to the outer boundary of the mass (and to the country rock structures in the aureole of deformation), but in some instances, a planar structure can be observed making an oblique intersection with the margin of an intrusion (Fig. 3). The planar structure tends to be best-developed in the marginal portions of the intrusion, where the rock may resemble an augen gneiss, but generally becomes less pronounced with increasing distance from the outer contact and the central portions of many

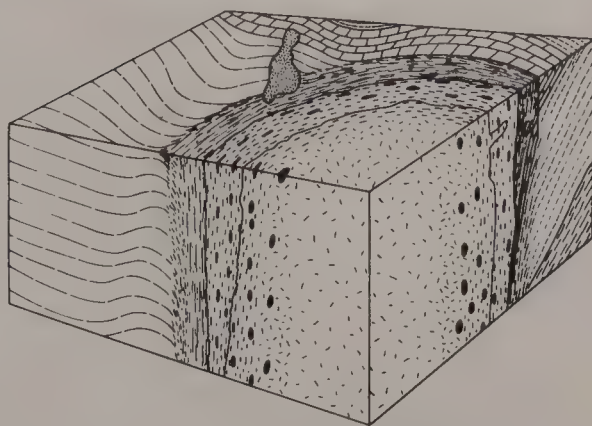


FIGURE 1. Generalized block-diagram of the northwestern part of the Ardara granitic intrusion, Donegal, Eire, showing deflection of the country rock structures and development of a zone of strong deformation in the vicinity of the intrusion. The planar structure within the intrusion is defined by oriented crystals and inclusions: this structure is almost lacking in the core of the mass but becomes more intense as the outer contact is approached, where the basic inclusions are not only well oriented, but are also much flattened. This planar structure continues across irregularities in the contact between the two units that form the mass, an inner granodiorite and an outer quartz monzodiorite. (From Pitcher and Berger, 1972.)

intrusions are virtually structureless. In some instances (Fig. 1) a planar structure crosses internal contacts between different units of an intrusive complex without deviation.

A lineation, defined by elongate, oriented elements (crystals, xenoliths) can sometimes be identified lying within the planar structure. H. Cloos, in particular (see references in Balk, 1937), has described intrusions where both planar and linear structures can be identified

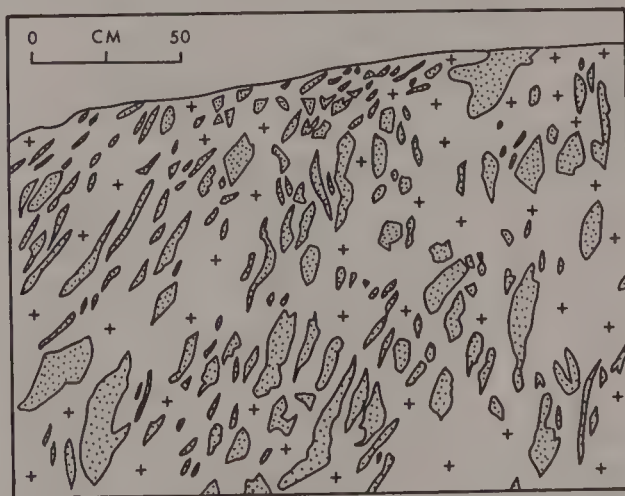


FIGURE 2. Aligned microdiorite xenoliths in the contaminated facies of the Puscao Unit, Coastal Batholith of central Peru. (After Cobbing and Pitcher, 1972.)

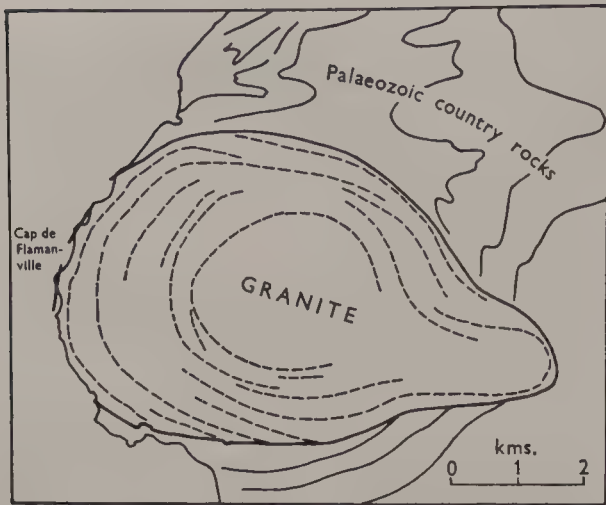


FIGURE 3. Steeply dipping planar structure that locally intersects the margin of the intrusion; Flammanville granite, Cotentin, France. (From Read and Watson, 1962.)

and where the planar structure defines a dome that contains an arch defined by the linear structure. Systems of fractures displaying a pattern of orientation that is related to the attitude of the planar and linear structures can be recognized in these intrusions (Fig. 4). Intrusions displaying such a variety of closely related primary structures appear to be rare. In most cases the orientation of the fracture systems appears to bear no simple relationship to that of the planar and linear structures, although there are systems of radial and tangential joints in certain intrusions.

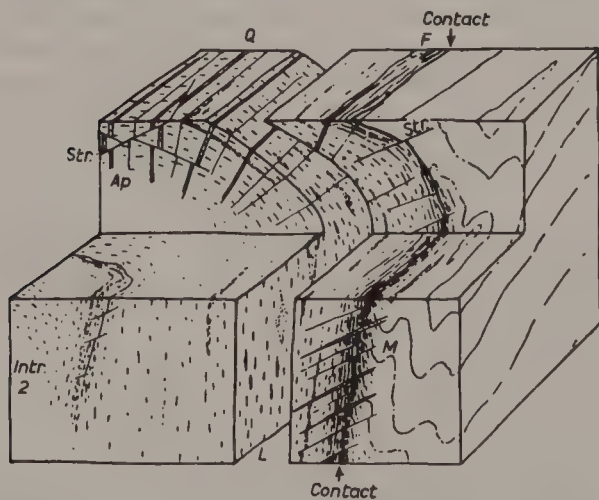


FIGURE 4. Block diagram of part of a batholith: M, marginal thrusts, some with injected aplite; F, planar structure; L, linear structure; Q, cross joints, some with injected aplite; Str, planes of stretching; Intr 2, second, concentric intrusion. Schistosity parallel to the granite contact is developed in the country rock and the axes of folds are tilted away from the intrusion. (From Hills, 1972.)

There seems to be general agreement that, in order to account for these structures, a mechanism can be invoked involving the peripheral distension of a mass of magma during emplacement. This distension, resembling the stretching of the skin of an inflating balloon, is thought to have affected the country rocks in the vicinity of the igneous mass as well as the marginal parts of the intrusion and to be due to the continued influx of magma into the center of the emplaced mass. There is some disagreement as to the state of the intruded material during at least the final stage of deformation, and, in particular, on the precise mechanism of production of the planar and linear structures in the igneous rocks. Cloos (and Balk) regarded these as having formed when at least a residue of silicate melt remained, but Berger and Pitcher (1970) argue that, as the planar structure sometimes transgresses the contact with the country rocks, then the country rocks and the marginal areas of the intruded mass must have deformed as a unit within which there was no significant difference in rigidity; they infer that deformation was continuing when the margin of the intrusion was essentially solid. It would seem that both explanations are probably valid, in some instances postconsolidation deformation being confined to fracture systems as described by Cloos, but in others, possibly at greater depth, involving the entire margin of the mass.

It should be noted that although the term “flow structure” is often used to describe the planar and linear structures in these intrusions, it is obviously inappropriate in the situation envisaged by Berger and Pitcher (1970). Even in the Cloos model, where the orientation of solid elements suspended in liquid is thought to occur, it is apparent that the distensional movements that must be regarded as being directed radially outwards are only reflected indirectly by the orientation of planar and linear elements in the igneous rock. It is possible that alignment of crystals and inclusions parallel to the direction of flow may occur when a nondistensional process is envisaged, e.g., in the conduit feeding the intrusion. However, it is difficult to visualize how this mechanism can produce a planar structure unaccompanied by a linear structure.

In some instances (see Bott et al. in Newall and Rast, 1970; Pitcher and Berger, 1972) geophysical data indicate that granitic intrusions extend to depths of the order of 10 km below the present surface. These data, in conjunction with the evidence reviewed above, lend support to the suggestion that the emplacement of a forcible intrusion may resemble the ascent of a salt dome; the term *diapiric* intrusion is

frequently used as a consequence. Granitic magma is envisaged as rising from depth, possibly simply because of its buoyancy, although the frequent association of such intrusions with orogeny suggests that regional compression may also promote intrusion (Balk, 1937; Pitcher and Berger, 1972). Experiments by Grout (1945; see Pitcher and Berger, 1972), suggest that the mass of ascending magma may acquire a tear-drop shape, possibly becoming detached from its root zone. The level of final emplacement will probably be determined by the rise in viscosity of the magma, which eventually overcomes the effect of the agencies promoting ascent.

The majority of readily identifiable forcible intrusions are granitic in character and have been emplaced in country rocks that have yielded by plastic deformation. It is possible that many other intrusions have been emplaced forcibly, but because of either the fluid, noncrystalline state of the magma or the brittle behavior of the country rocks, no record of the emplacement mechanism has been preserved. Thus, it is possible that the emplacement of many basic sills involves the forceful displacement (uplift) of the surrounding rocks and, at high level, doming of a thin carapace of country rocks accomplished largely by brittle fracturing and faulting that may not be readily recognizable (Fig. 5).

Passive (permitted) intrusions. These are intrusions that have been emplaced by processes of displacement of the country rocks that do not appear to be dominated by outwardly directed pressures from the magma, i.e., in contrast to forcible intrusions, where the pressure within the magma must be envisaged as being greater than in the surrounding rocks and where the magma viscosity seems generally to be high, in passive intrusions the relative fluidity of the magma seems to be the dominant feature. In some areas these intrusions follow earlier forcible intrusions.

The process of emplacement often seems to involve the exploitation of fractures in the country rocks, space for the magma apparently being created by the detachment of blocks of country rock that then sink into the magma. This process is called *stopping* and, although it is envisaged that large intrusions may be emplaced by this mechanism, the actual stopping process is generally thought to operate on a small scale (piecemeal) and to involve the detachment of blocks with dimensions of only a few tens or hundreds of meters (Fig. 6). Evidence of stopping on a grander scale is provided by intrusions where huge cylindrical blocks of country rocks have clearly subsided into a body of magma, producing intrusions with an annular

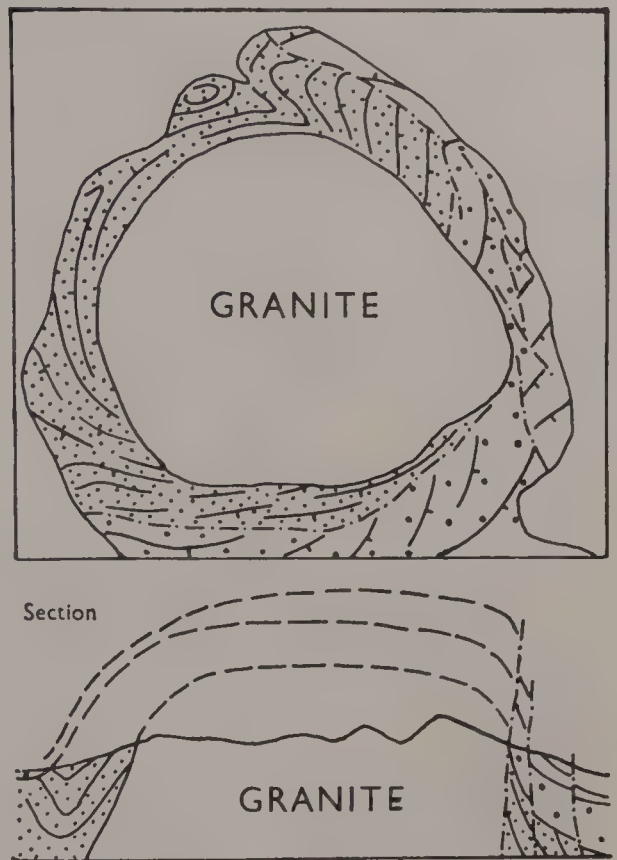


FIGURE 5. Faults and folds produced by the emplacement of the Arran granite, Scotland. (From Read and Watson, 1962.)

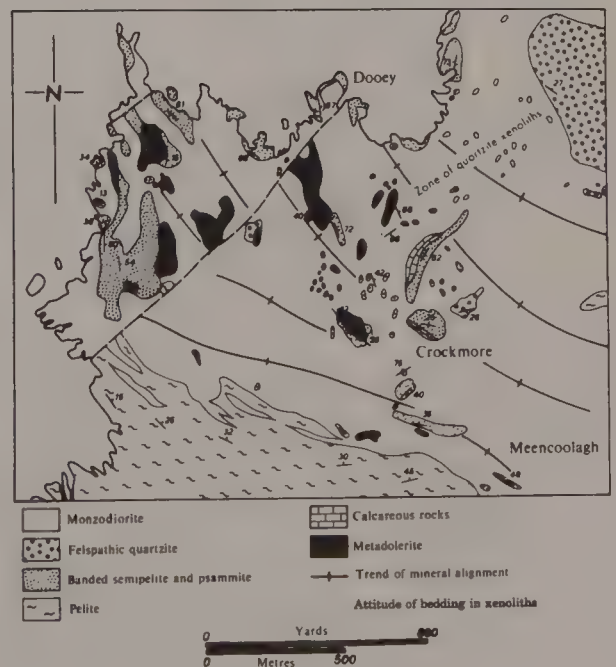


FIGURE 6. Part of the Fanad intrusion, Donegal, Eire, showing evidence of piecemeal stopping and the preservation of "ghost stratigraphy." (From Pitcher and Berger, 1972.)

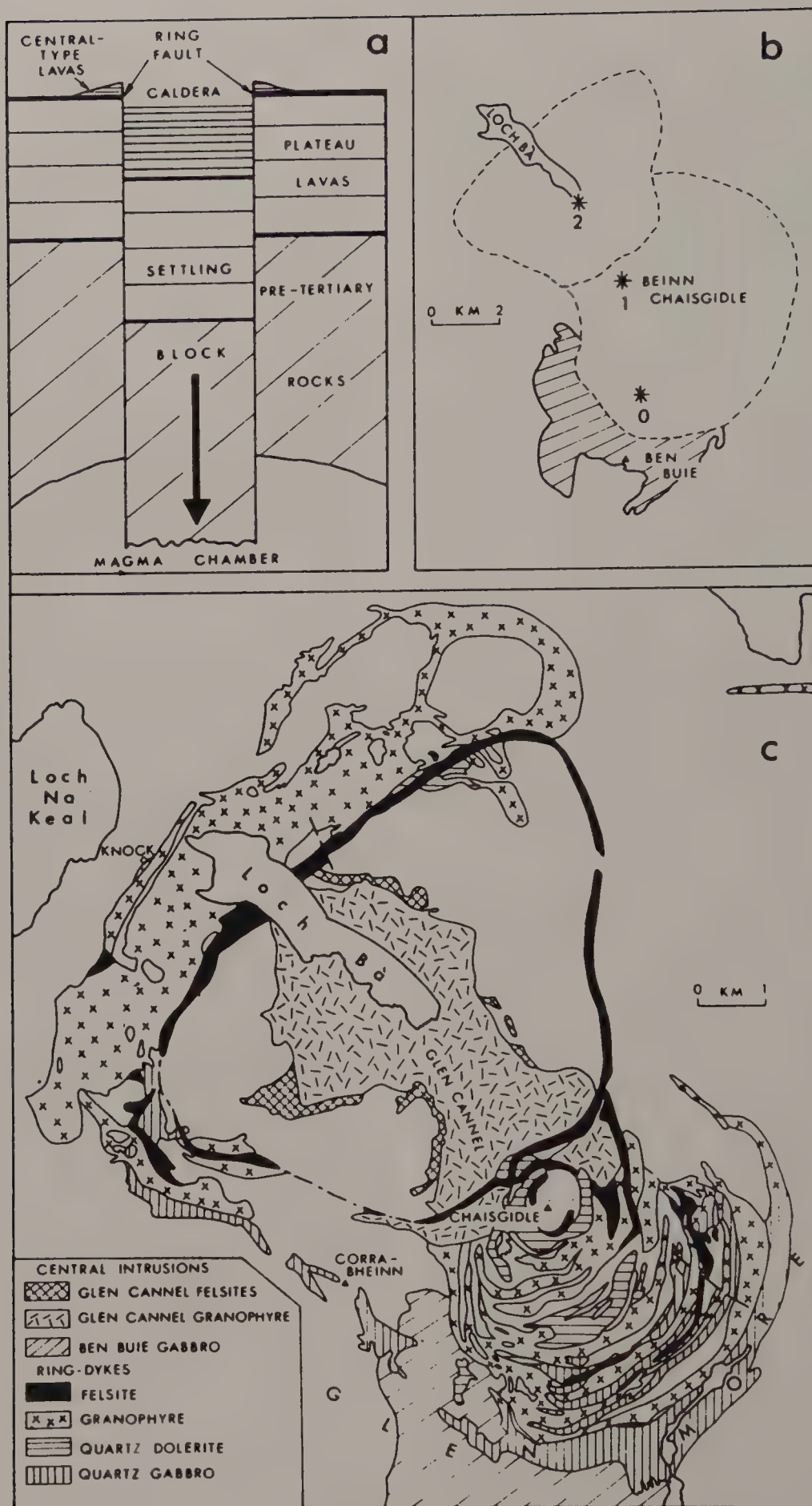


FIGURE 7. Ring dykes and major intrusions in central Mull, Scotland (c) associated with cauldron subsidence (a) and the successive emplacement of intrusions about centers (*) 0, 1, 2 (b). (From Donaldson, 1983.)

outcrop (*ring dykes*—Fig. 7). This process is termed cauldron subsidence and when found in association with volcanicity in high-level intrusions is thought to be accompanied by caldera formation at the surface. However, in other instances there would seem to be no need to relate the foundering of these huge blocks to surface events and the formation of “bell-jar” intrusions by processes (reduction of magma pressure is thought to be particularly important) involving the production of inward-curving fractures has been postulated (see Roberts, in Newall and Rast, 1970).

The term permitted (or permissive) intrusion is sometimes used in relation to these intrusive masses (e.g., Pitcher and Berger, 1972) but as this term had already been used by Mayo (cited in King, 1954) for intrusions that were thought to have been emplaced into areas of relatively low pressure produced by regional deformation, continued use of this term would seem to be undesirable.

It should be noted that evidence of piecemeal stoping is generally limited but may be provided by “ghost stratigraphy,” i.e., the distribution of scattered country rock xenoliths sometimes appears to preserve original stratigraphic relations (Fig. 6). Stopping is generally associated with acidic intrusions but has also been invoked to account for the emplacement of basic intrusions. Problems arise with the latter, however, particularly when cauldron subsidence is envisaged, as the density of the country rocks may be no greater than that of the magma.

M. MUNRO

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Cross-references: *Banding in igneous rocks; Batholith; Calderas and volcano-tectonic depressions; Cone sheets; Contact (thermal) metamorphism; Diapirism; Dyke (dike); Emplacement—modes; Granites and volcanism; Intrusion; Migmatite—structural relationships; Ring dyke; Sill.*

IGNEOUS PETROLOGY

Igneous petrology relates to the study of rocks that solidified from hot, molten or partly molten material, i.e., *igneous rocks* (Latin *ignis*, fire). It includes *volcanology* (*vulcanology*) that deals with the processes by which magma and its gases rise into the crust and are extruded on the Earth's surface and into the atmosphere (*volcanism*), and with the nature of the rocks formed by these processes. The description and systematic classification of such igneous rocks is referred to as *igneous petrography*, the study of the chemical composition of such rocks as *igneous petrochemistry*, and the elucidation of the origin and formation of these rocks as *igneous petrogenesis*.

D. R. BOWES

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Cross-references: *Igneous petrology—terms; Igneous rocks—classification and nomenclature; Igneous rocks—tectonic setting; Petrology—branches; Petrology—history; Volcanoes and volcanism.*

IGNEOUS PETROLOGY—TERMS

Chemical aspects

Acid(ic)—see **Acid(ic) igneous rocks**.

Alkaline (alkali, alkalic)—refers to a rock with (1) more alkali metals than is considered average for the group of rocks to which it belongs, (2) more Na and K than is required to form feldspar with the available Si, or (3) an alkali–lime index <51, or a rock belonging to the Atlantic suite.

Alkali–lime index (alkali–calc index)—based on the wt.% of SiO_2 present when wt.% of CaO and $\text{K}_2\text{O} + \text{Na}_2\text{O}$ are equal; four chemical classes of igneous rocks are recognized: *alkalic* (when $\text{SiO}_2\%$ < 51), *alkali–calcic* (51–56), *calc–alkalic* (56–61), and *calcic* (>61).

Basic—see **Basic igneous rocks**.

Critically undersaturated—refers to a rock having feldspathoids and olivine but no hypersthene in the norm.

Intermediate—see **Intermediate igneous rocks**.

Mediosilicic—a term equivalent to intermediate.

Metaluminous—refers to a rock in which the molecular proportion of Al_2O_3 is > that of Na_2O and K_2O combined but generally < that of $\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO}$.

Oversaturated—refers to (1) a rock containing sufficient SiO_2 for free silica (quartz, tridymite, etc.) to form; (2) a magma from which such a rock crystallizes.

Peralkaline—refers to a rock with the molecular proportion of Al_2O_3 < that of $\text{Na}_2\text{O} + \text{K}_2\text{O}$.

Peraluminous—refers to a rock with the molecular proportion of Al_2O_3 > that of $\text{Na}_2\text{O} + \text{K}_2\text{O}$.

Persilicic—a term equivalent to acid(ic).

Saturated—refers to (1) a mineral that can form in the presence of free SiO_2 ; (2) a rock mainly composed of these minerals; (3) a rock having quartz in its norm; (4) a magma from which (2) and (3) form.

Silica coefficient—the ratio of total SiO_2 in a rock to silica in the feldspars and metasilicates.

Silicic—a SiO_2 -rich rock or magma (generally taken as being > 65% SiO_2); in addition to silica

combined as feldspars, etc., free SiO_2 is expressed as quartz, tridymite, etc., as in granite and rhyolite; *acid(ic)* is a synonymous term.

Subalkalic—refers to (1) a rock containing no alkali minerals other than feldspars; (2) a rock of the Pacific suite.

Subaluminous—refers to a rock in which there is little or no excess of Al_2O_3 over that required to form feldspars or feldspathoids.

Subsilicic—a term equivalent to basic.

Ultrabasic—see **Ultrabasic igneous rocks**.

Undersaturated—refers to (1) rocks containing unsaturated minerals (e.g., feldspathoids or olivine); (2) rocks containing unsaturated minerals in the norm; (3) a magma from which such rocks crystallize.

Unsaturated—refers to a mineral that does not form in the presence of free silica (e.g., feldspathoids, olivine).

Magmatic aspects

Connate—refers to fluids derived from the same magma, e.g., water entrapped in an effusive igneous rock at the time of formation.

Cotectic—refers to conditions of temperature, pressure, and composition under which two or more solid phases crystallize together (without resorption) from a single liquid over a particular range of falling temperature; the line or surface representing the corresponding phase boundary on the liquidus of a phase diagram is referred to as a cotectic line or cotectic surface.

Critical pressure—the pressure needed to condense a gas at the critical temperature above which, regardless of pressure, the gas cannot be liquified.

Critical temperature—the temperature above which a substance can only exist in a gaseous state, regardless of the pressure exerted.

Decadent stages—those late stages of volcanism during which solfataras, fumaroles, geysers, and hot springs develop.

Gaseous transfer—separation, in a magma, of a gaseous phase that moves relative to a magma and releases dissolved substances.

Hydrothermal stage—that stage in the cooling of a magma during which the residual fluid is greatly enriched in water and other volatile constituents; the temperature is lower than for the pegmatitic stage (see **Hydrothermal processes**).

Hyperfusible—refers to any substance that is capable of lowering the melting range of an end-stage magmatic fluid.

Liquidus—the locus of points in a temperature–composition diagram that represents the maximum solubility (saturation) of a solid component or phase in a liquid phase; at

temperatures above the liquidus the system is completely liquid.

Orthomagmatic stage—the stage during which as much as 90% of the crystallization of silicates from a magma takes place.

Outgassing—the removal from magma of occluded gases, particularly water vapor.

Pegmatitic stage—that stage in the cooling of a magma at which the residual fluid is sufficiently enriched in volatile materials to permit the formation of pegmatites (see **Pegmatite**).

Pneumotectic—refers to processes and products of magmatic crystallization affected to some degree by gaseous constituents of the magma.

Pneumatolytic stage—that stage in the cooling of a magma during which the solid and gaseous phases are in equilibrium (see **Pneumatolysis**).

Postmagmatic (stage)—refers to reactions or events that occur after the crystallization of the bulk of the magma, and usually including the hydrothermal stage.

Reaction boundary (curve, line)—the position at which a crystalline phase in a system reacts with the liquid as temperature declines to form another crystalline phase.

Residual liquid—the mainly volatile components of a magma that remain in a magma chamber after much of the crystallization has taken place.

Solidus—the locus of points in a system on a temperature–composition diagram at temperatures above which solid and liquid are in equilibrium and below which the system is completely solid.

Subsolvus—refers to those granites, syenites, and nepheline syenites that are characterized by both K-feldspar and plagioclase.

Supercooling—the process by which the temperature of a phase or assemblage is lowered below the temperature (or temperature range) at which a phase change should occur at equilibrium; it generally refers to a liquid taken below its liquidus temperature.

Supercritical—refers to a system at a temperature greater than its critical temperature.

Superheating—(1) the addition of more heat than necessary to complete a given phase change; (2) the addition of more heat than necessary to cause complete melting; (3) the process of increasing heat beyond the point at which a phase or assemblage changes at equilibrium.

Mineralogical aspects

Corrosion—The process of modification of the outer parts of early-formed crystals or xenoliths by the host magma as a result of partial resorption, dissolution, or fusion; second-

dary minerals around the original material constitute a *corrosion border* (rim) or *corona*.

Dolimorphic—refers to an igneous rock in which released minerals are prominent, e.g., lamprophyre with biotite and quartz but little hornblende.

Heteromorphic—refers to igneous rocks having similar chemical composition but different mineralogical composition.

Occult mineral—a mineral whose actual or potential presence in an igneous rock is deduced from chemical considerations rather than by petrographic methods.

Rosenbusch's law—a statement of the sequence of crystallization of minerals from magmas proposed by H. Rosenbusch in 1882; subsequent modifications have been made.

Other terms

Apomagmatic—refers to a hydrothermal mineral deposit situated at an intermediate distance from its magmatic source.

Clan, class, family, order—terms used in rock classifications for groups of igneous rocks that are closely related in chemical composition.

Comagmatic—refers to an igneous suite having a common set of chemical, mineralogical, and textural features indicative of derivation from a common parent magma.

Cryptomagmatic—refers to a hydrothermal mineral deposit without demonstrable relationship to an igneous mass.

Crystallization magnetization—a stable remnant magnetization resulting from slow growth during crystallization of magnetically ordered mineral grains in a magnetic field.

Effusive—refers to those rocks that crystallized from magma (lava) poured out from a vent or fissure.

Endogene effects—the effects of an igneous mass on the margin of the mass itself.

Endogenic (endogenetic)—refers to the processes, and resultant products, that originate within the Earth, e.g., volcanism and volcanoes.

Eruptive—refers to rocks formed by solidification of magma; sometimes restricted to effusive rocks.

Extrusive—refers to igneous material that has been ejected on to the Earth's surface, both effusive material (lava flows) and *ejectamenta* (*pyroclastic deposits*); the products of extrusion are extrusive rocks.

Intratelluric—refers to (1) a phenocryst of an earlier generation than its groundmass that formed at depth prior to the upward movement of the magma; (2) the period of magmatic crystallization at depth just prior to effusion of the magma as lava; and (3) material located, formed, or originating at depth.

Palaeotypal—refers to aphanatic and porphyritic rocks with characteristics of altered effusive or hypabyssal rocks, such as many rocks of pre-Tertiary age.

Perimagmatic—refers to a hydrothermal mineral deposit situated near to its magmatic source.

Pyrogenesis—the intrusion and extrusion of magma and its derivatives.

Telemagmatic—refers to a hydrothermal mineral deposit situated far from its magmatic source.

Volcanogenic—of volcanic derivation; commonly used for mixed effusive and pyroclastic assemblages.

D. R. BOWES

IGNEOUS ROCKS—CLASSIFICATION AND NOMENCLATURE

More than 1,000 igneous rocks have been named according to their characteristics determined in field, microscopic, chemical, and genetic studies. As the igneous rock nomenclature has expanded, schemes of classification have been constructed (1) to ensure that all scientists refer to a given rock type by the same name, (2) to enable geologists to compare one igneous rock with another, and (3) to provide a means for thinking about rocks in groups.

Methods of classification

Igneous rocks have been classified on the basis of the following fundamental characteristics: mineralogical composition, chemical composition, texture, structure, color index, age, mode of occurrence, and genesis. Various combinations of these characteristics have been selected for each classification, depending upon the data available and the purpose of the classification.

The early classifications, such as those used for megascopic purposes today, were simple field classifications, based upon megascopic characteristics, usually mineralogy and texture (size, shape, and arrangement of mineral grains). With the advent of the application of the microscope to the study of rocks, beginning in the latter part of the 19th century, igneous rocks were found to contain a much greater mineralogical and textural variety than had been detected megascopically. These laboratory studies with the microscope led to more elaborate classifications based upon texture and qualitative or quantitative determinations of the amounts of each mineral present (e.g., classification of Johannsen, 1931–1938). Also,

during the 19th century many chemical analyses of igneous rocks provided the bases for chemical classifications, such as the CIPW system that appeared in 1902. Igneous rocks may also be grouped together, for purposes of discussion, into those that are closely associated in time and space owing to linkage through a common genesis. Although the use of more than one system of classification causes some problems of correlation from one system to another, the advantages of the availability of a classification for each type or combination of fundamental characteristics exhibited by igneous rocks appears to outweigh the disadvantages.

Field classifications. Simplified classifications, such as that shown in Table 1, are utilized in the field to group igneous rocks according to those features that can be determined with the eye. The fundamental characteristics used as parameters for field classifications are normally texture and mineralogical composition.

The textural aspects most important include the degree of crystallinity (entirely crystallized or holocrystalline, partly crystallized, or merocrystalline (hypocrystalline), entirely uncrystallized as a glass, or holohyaline) and the average grain size. (Here fine, medium, and coarse-grained are taken to be <1 mm, 1–5 mm, and >5 mm, respectively; other authors use <0.01 (0.05) mm, 0.01(0.05)–1(2) mm, and >1(2) mm). Fine-grained rocks are called *aphanitic*; medium- and coarse-grained ones are termed *phaneritic*. Fine-grained rocks occur primarily as lavas, medium-grained rocks occur as minor intrusions (e.g., dykes, sills), while coarse-grained rocks are most common as large bodies that crystallized slowly at depth.

Igneous rocks that have formed by cooling at a constant rate tend to develop a uniform grain size (equigranular), whereas those that experience an abrupt change in their rate of cooling commonly form crystals of two distinctly different sizes (porphyritic). Table 1 gives names for igneous rocks that are essentially uniform in grain size. For rocks in which two distinctly different grain sizes are present, the names shown in Table 1 are followed by the term “porphyry” (e.g., rhyolite porphyry, or porphyritic rhyolite). A problem arises for rocks in which large grains (phenocrysts) constitute >50%. Here they are named according to their coarse-grained equivalents followed by the term “porphyry”: an aphanitic porphyry consisting of orthoclase and quartz, in which the phenocrysts comprise >50% of the rock, is called a granite porphyry (in other classifications it would be a porphyritic rhyolite), while a porphyritic granite consists of <50% phenocrysts in a coarse-grained matrix.

TABLE 1. Field classification of igneous rocks

Glassy	Obsidian Pitchstone Perlite						
Fine-grained aphanitic	Felsite					Trap	
	Rhyolite	Trachyte	Latite (Trachyandesite)	Dacite	Andesite	Basalt	
Medium-grained phaneritic	Aplite					Dolerite Diabase	
Coarse-grained, phaneritic	Granite	Syenite	Monzonite	Quartz diorite	Diorite	Gabbro	Peridotite Dunite Pyroxenite Hornblende
	0–10 C.I.		10–40 C.I.			40–70 C.I.	70–100 C.I.
	Qz	No Qz		Qz	No Qz		
				Na Pl		Ca Pl	
	Kf > Pl		Kf = Pl	Pl > Kf			No feldspar

Mineral composition forms the other parameter of Table 1. The *essential constituents* (those necessary for the classification of a rock) are shown below each rock type; e.g., K-feldspar and quartz are the essential minerals in a granite. *Characterizing accessory minerals* are not necessary to the classification of a particular igneous rock, but they may be used to precede a rock name; e.g., a granite with a moderate proportion of hornblende may be called hornblende granite, and if the rock also contains appreciable biotite (in excess of hornblende) it is called a biotite–hornblende granite. Minor *accessory minerals* are those not utilized in the igneous rock names.

A close relationship exists between the mineralogical composition of an igneous rock and its chemical composition. Acidic igneous rocks (>66% SiO₂) tend to consist primarily of quartz and K-feldspar and they are positioned toward the left-hand portion of Table 1 (e.g., granite). Intermediate rock types (52–66% SiO₂) contain both K-feldspar and plagioclase, and occupy intermediate positions in the field classification (e.g., monzonite). Basic rocks (45–52% SiO₂), such as gabbro, consist primarily of calcic plagioclase. Ultrabasic rocks (<45% SiO₂) consist of femic minerals with little or no feldspar and occur in the right-hand column of Table 1.

The mineral-based terms mafic and ultramafic are close to, but not exactly synonymous with, the above chemically-based terms basic and

ultrabasic. The former terms are based upon the ratios of femic (or dark-colored) to salic (or light-colored) minerals in a rock, expressed as a color index (C.I.). In mafic rocks C.I. is 40–70, while in ultramafic rocks it is >70. Rocks deviating significantly from their common C.I. are prefixed by leuco- or mela-; e.g., a syenite with very abundant dark minerals is called a melasyenite. Since the dark minerals normally have specific gravity >2.8 and light ones <2.8, the rocks generally become progressively heavier from left to right in Table 1.

A root name as shown in Table 1 may be preceded by the color, texture, alteration, and characterizing accessory minerals; e.g., a complete field name might be a pink, coarse-grained, equigranular, kaolinized, biotite–hornblende granite.

It is important to remember that the names given to igneous rocks utilizing a field classification indicate *only* their megascopic characters. Additional field classifications are given by Brown (1952), Travis (1955), and Streckeisen (1973) and many other authors.

Laboratory classifications. The *petrographic classification* of Johannsen (1931) is one of the best-known quantitative modal classifications: classes, orders, and families based, respectively, on percentage of dark minerals, plagioclase feldspar composition, and the quartz–orthoclase–plagioclase–nepheline (plus other silica-poor feldspathoids) ratios have been erected. Each rock name has an equivalent

TABLE 2. Igneous rock nomenclature chart designed to enable the user to find the meaning of a given name, not to find a name for a given specimen

		Kf 90–100 Pl 0–10	Kf 67–90 Pl 10–33	Kf 33–67 Pl 33–67	Kf 10–33 Pl 67–90	Kf 0–10 Pl 90–100	Little or no feldspar	
					An 0–50	An 50–100		
Quartz over 5%	Aphanitic	RHYOLITE (LIPARITE) Aporhyolite (Tex) QZ PORPHYRY 1	QZ LAITITE (RHYODACITE) (DELLENITE) Toscanite (Kf) 2	DACITE 3	Qz basalt THOLEIITE (norm Qz) 4			
	Medium-grained, phaneritic				QZ DIABASE (QZ DOLERITE) 7			
	Coarse-grained, phaneritic	APLITE → Granophyre → 6	GRANITE Luxullianite (Tu) Rapakivi (Tex) Charnockite (Hy) → 9	GRANODIORITE Opdalite (Hy) Qz mangerite (Per, Cpx) 11	QZ DIORITE (TONALITE) Trondjemite (Qz) 12	Qz gabbro 13		KOMATIITE 17a
Quartz less than 5%	Aphanitic	TRACHYTE Domite (Olig) Vulsinite (Ca Pl) Ciminite (Ol) 14		LATITE (TRACHYANDESITE) Tahitiite (Hn) 15	ANDESITE PROPYLITE (altered) Weisbergite (Aug) Trachybasalt 16	BASALT OL BASALT (Ol) PICRITE BASALT (Ol) (OCEANITE) Ankaramite (Ol, Px) Oligoclase basalt (Olig, Kf) Mugearite (Tex) Spilitite (Ab) 17		
	Medium-grained phaneritic	Sölvbergite Bostonite (Tex) Minette (Lam, Bio) Vogesite (Lam, Hb, Aug) → 18	Perthosite 19	Albitite (Ab) Kersantite (Lam, Bio) Spessartite (Lam, Hb, Aug) → Camptonite (Lam, Alk) 20	DIABASE (DOLERITE) Oudinite (Lam, Aug, Hb) 21	PERIDOTITE LHERZOLITE (Opx = Cpx) HARZBURGITE (Opx) (SAXONITE)		
	Coarse-grained, phaneritic	ALKALI SYENITE Nordmarkite (Qz) Fenite (Aeg) Larvikite (Tex) → 18	SYENITE QZ SYENITE (Oz) Shonkinite (FM) 20	DIORITE Mangerite (Per, Hy) ← Appinitite (FM) 21	GABBRO HYPERITE (Opx = Cpx) NORITE (Opx) OL GABBRO (Ol)			

TABLE 2. (continued)

Alnöite	36	Gabbro	27	Melteigite	41	Pitchstone	*	Theralite	40
Analcime syenite	34	Glenmuirite	35	Miaskite	37	Propylite	16	Tholeiite	4
Analcimite	33	Granite	9	Minette	18	Pulaskite	37	Tinguaite	34
Andesite	16	Granodiorite	11	Monchiquite	36	Pumice	*	Tonalite	12
Ankaramite	17	Granophyre	6	Monmouthite	41	Pyroxenite	28	Toscanite	2
Anorthosite	27	Groandite	5	Monzonite	24	Quartz basalt	4	Trachyandesite	15
Aplite	5,6	Harzburgite	28	Mugearite	17	Quartz bostonite	5	Trachybasalt	16
Aporhyolite	1	Hornblendite	28	Murambite	32	Quartz diorite	7	Trachyte	14
Appinite	26	Hyperite	27	Nepheline andesite	31	Quartz dolerite	12	Troctolite	27
Basalt	17	Ijolite	41	Nepheline latite	30	Quartz gabbro	7	Trondhjemite	12
Basanite	32	Jacupirangite	41	Nepheline monzonite	38	Quartz keratophyre	13	Tuff	*
Bostonite	18	Kaiweike	14	Nepheline syenite	37	Quartz latite	1	Ugandite	33
Camptonite	20	Katungite	33	Nepheline syenodiorite	39	Quartz mangerite	2	Umpiteke	37
Cedricite	33	Kentallinite	24	Nepheline	33	Quartz monzonite	11	Uncompagrite	41
Charnockite	9	Kenyte	29	Nordmarkite	22	Quartz porphyry	10	Urtite	41
Ciminite	14	Keratophyre	14	Norite	27	Quartz syenite	1	Vitrophyre	*
Comendite	1	Kersantite	20	Obsidian	*	Quartz syenite	23	Vogesite	18
Cortlandtite	28	Kimberlite	28	Oceanite	17	Raplanite	40	Vulsinite	14
Craigmontite	40	Kivite	32	Odinite	21	Rhyolite	9	Websterite	28
Crininite	35	Komatite	17a	Oligoclase basalt	17	Saxonite	2	Wehrhite	28
Dacite	3	Kyllite	40	Olivine basalt	17	Schriesheimite	1	Weiselbergite	16
Dellenite	2	Lardalite	38	Olivine gabbro	27	Scoria	28	Welded tuff	*
Diabase	21	Larvikite	22	Opdalite	11	Seyelite	*	Wolgidite	33
Diorite	26	Latite	15	Orendite	29	Shonkinite	23		
Ditroite	37	Leucite	33	Ouachitite	36				
Dolerite	21	Lherzolite	28	Paisanite	5				

* Rocks whose names are based mostly upon their textures and whose compositions may range widely are listed here but they are not shown on the chart.

Ab	Albite	Di	Diopside	Lc	Leucite	Phl	Phlogopite
Aeg	Aegirine	Fe	Fe-rich rock	Leuco	Leucocratic	Pl	Plagioclase
Alk	Alkali-rich rock	FM	Ferromagnesian mineral abundant, usually > 50%	Me	Melilitite	Px	Pyroxene
Alt	Altered	Hb	Hornblende	Mela	Melanocratic	Qz	Quartz
An	Anorthite	Hs	Hastingsite	Na	Na-rich rock	Rie	Riebeckite
Anal	Analcime	Hn	Häuyne	Na Amp	Na-amphibole	Sd	Sodalite
Aug	Augite	Hy	Hypersthene	Ne	Nepheline	Tex	Texture
Bio	Biotite	Int Pl	Intermediate plagioclase	Norm	Normative	Ti Bio	Titaniferous biotite
By	Bytownite	K	K-rich rock	Ol	Olivine	Tu	Tourmaline
C	Corundum	Kf	K-feldspar	Olig	Oligoclase		
Ca Pl	Calcic plagioclase	Lab	Labradorite	Opx	Orthopyroxene		
Cpx	Clinopyroxene	Lam	Lamprophyric texture	Per	Perthite		

three-digit number symbol, the first digit indicating the class, the second for order, the third for family, followed by the letter P or E for plutonic (intrusive) or extrusive. For example, the symbol for a normal granite is 226P.

Streckeisen (1967) has proposed a petrographic classification, similar in many respects to that of Johannsen, with an initial subdivision into two classes based upon their dark-mineral content. The largest class, those with <90% dark minerals, are named according to their proportions of quartz–K-feldspar (plus albite)–plagioclase–feldspathoids. Leuco- and mela- are used as prefixes for specimens lighter and darker than normal. Rocks containing >90% dark minerals are named according to the proportion of those dark minerals. This classification for the plutonic igneous rocks has been modified by the International Union of Geological Sciences Subcommittee on the Systematics of Igneous Rocks (Streckeisen, 1967, 1973, 1976). Additional efforts by the Subcommittee have resulted in the presentation of similar classifications for volcanic rocks, lamprophyres, carbonatites, and melilitic rocks (Streckeisen, 1978, 1979) and for pyroclastic deposits and fragments (Schmid, 1981) (see also Le Maitre, 1989).

Other petrographic classifications include those of Shand (1949), Rittmann (1952), O'Connor (1965), and Hatch et al. (1972).

Chemical classifications of igneous rocks are based upon major element analytical data or calculation of these into theoretical standard (normative) minerals (see **Petrochemical calculations** for information about CIPW norms, molecular norms, Niggli numbers, etc.; see also **Differentiation index**).

Two widely used classifications for certain igneous rocks are those that include aspects of both normative and modal systems. O'Connor's (1965) classification for quartz-rich (>10% quartz) rocks emphasizes a direct comparison of the normative and modal compositions of plutonic rocks and provides a normative classification for the volcanic rocks of those compositions. Rittman's (1952) classification of volcanic rocks involves three keys and corresponding diagrams for rocks whose mode has been measured or estimated and whose chemical composition has been determined. A graphical procedure for the last classification eliminates some of the calculations inherent in normative systems. These and other calculations are now done routinely by computer.

Further use of chemical data as a basis for classification is given by Streckeisen and Le Maitre (1979) and De la Roche et al. (1980).

Genetic classifications. These depend on

interpretation and reflect ideas prevalent at the time, as illustrated by Shand (1949), Tyrrell (1958), Carmichael et al. (1974) and Cox et al. (1979).

Nomenclature

With the growth of the number of igneous rock terms from <500 to >1,000 in the first half of the 20th century (e.g., see Sørensen, 1974) for a comprehensive glossary of names and definitions of alkaline rocks), igneous petrologists should consider a system of more self-explanatory rock nomenclature, somewhat similar to that of metamorphic petrologists, e.g., nepheline dolerite for crinanite, analcime–nepheline dolerite for teschenite, etc. In the meantime, Table 2 (adapted from Hagni, 1965) sets out a simple tabular guide to the approximate mineralogical and textural characteristics of a large number of the igneous rock terms most commonly used in the literature. It is similar in outline to the field classification in Table 1, and enables one to determine quickly that an igneous rock name encountered in the literature is similar to some common rock and to find out the principal differences that set that rock aside from the more common one. The approximate abundance of the rocks shown in Table 2 is indicated by their capitalization. A term that is synonymous, or nearly so, with another term is placed below that term and within parentheses. Thus “(tonalite)” is positioned below “quartz diorite.” Mineral abbreviations, placed within parentheses and following rock terms, indicate the principal mineralogical difference between that rock and the preceding igneous rock or from the most common rock in that compartment. Underlined and doubly underlined abbreviations indicate that mineral to be more abundant than in the preceding rock. The mineral abbreviation is preceded by “only” when that mineral is absent or nearly absent from that rock. An element is placed in parentheses after a rock that is characterized by an unusual abundance of that element. “(Tex)” follows a rock whose principal difference from the preceding rock lies in its textural aspects; “(Lam)” indicates lamprophyric texture. The accompanying alphabetical list of rock terms indicates those terms shown and their locations; each compartment is numbered to facilitate the location of a given term.

RICHARD D. HAGNI

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Cross-references: *Acid(ic) igneous rocks*; *Alkaline rocks—undersaturated*; *Basic igneous rocks*; *Calc-alkaline rocks*; *Intermediate igneous rocks*; *Peralkaline igneous rocks*; *Petrochemical calculations*; *Plutonic rocks—classification and nomenclature*; *Ultrabasic igneous rocks*; *Ultramafic rocks*.

IGNEOUS ROCKS—OTHER TERMS

The following are terms used for igneous rocks. They are additional to those set out in **Igneous rocks—classification and nomenclature** and under various individual headings such as **Basalt**, **Carbonatite**, **Gabbro** and **norite**, **Granite**, **Lamprophyre**, etc.; most are little used and many have strong local connotations. Many references to original definitions are given by Holmes (1928).

Aegipite—a calcite-apatite-aegirine-augite rock.

Allochetite—a porphyritic, hypabyssal rock composed of labradorite, orthoclase, titanite, nepheline, and orthoclase (Allochet Valley, Italy).

Ampasimenite—an ijolite porphyry or nepheline porphyry with glass in the matrix (Ampasimena, Malagasy).

Anegite—contains pyroxene, spinel, pyrope, and hornblende; feldspar and olivine are absent.

Anorthoclase—almost entirely composed of anorthoclase.

Apanteite—a nepheline-bearing apatite.

Arizonite—a hypabyssal dyke rock composed mainly of quartz and orthoclase (Arizona, U.S.A.).

Atatschite—a porphyritic rock with phenocrysts of orthoclase, augite, and biotite in a glassy groundmass containing small proportions of sillimanite and cordierite (Atatsch Mt., Urals, U.S.S.R.).

Baldite—a hypabyssal porphyritic rock composed of pyroxene phenocrysts in a groundmass of analcime, augite, and opaque ores; equivalent of analcime-bearing basalt.

Barshawite—an andesine-orthoclase lugalite.

Beschtauite—a sodic quartz porphyry (quartz keratophyre) (Mt. Beschtau, Caucasus Mtns., U.S.S.R.).

Cavalorite—a granular plutonic rock composed almost entirely of feldspar, with orthoclase more abundant than oligoclase.

Columbretite—an effusive rock composed of sanidine and altered hornblende laths in a dense groundmass of corroded oligoclase microlites with interstitial sanidine enclosing leucite grains.

Dahamite—an albite-rich paisanite or microgranite (Dahamis, Socotra Island, Yemen).

Damkjernite—a porphyritic hypabyssal rock with phenocrysts of biotite and pyroxene in a fine-grained groundmass of pyroxene, biotite, and magnetite with interstitial, nepheline, microcline, and calcite (possibly primary).

Euktolite—see *Venzanite*.

Eurite—a hypabyssal rock of granitic composition; a microgranite.

Farrisite—an aphanitic rock mainly composed of diopside and barkevikite with smaller

proportions of lepidomelane, olivine, and magnetite; feldspar and nepheline are mostly altered to, or replaced by, zeolites (Farris See, Norway).

Felsite—a leucocratic hypabyssal or effusive rock composed mainly of feldspar and quartz; quartz porphyry is a more commonly used term for porphyritic varieties.

Gabbrophyre—a porphyritic hypabyssal rock with phenocrysts of labradorite and augite in a groundmass of labradorite and hornblende.

Granolite—a plutonic rock characterized by granitic rather than porphyritic texture.

Helsinkiite—a hypabyssal rock mainly composed of albite and epidote with a hypidiomorphic-granular texture (Helsinki, Finland).

Heronite—a hypabyssal rock composed of spheroidal groups of orthoclase phenocrysts in a groundmass of radiating bundles of labradorite and acmite crystals with interstitial analcime; an altered tinguaitite (Heron Bay, Ontario, Canada).

Heumite—a dark-colored hypabyssal rock composed of alkali feldspar, barkevikite, and smaller proportions of nepheline, sodalite, and diopside (Heum, Norway).

Highwoodite—a dark-colored intrusive rock composed of loxoclase, labradorite, pyroxene, biotite, opaque ore, apatite, and possibly a small proportion of nepheline (Highwood Mtns., Montana, U.S.A.).

Hilairite—a porphyritic rock with phenocrysts of albite, nepheline, sodalite, acmite, and eudialyte in a trachytic groundmass.

Hirnantite—a variety of albite keratophyre or albitized tholeiite (Hirnant, N Wales).

Hollaite—a rock produced by interaction of sövite and rocks of the ijolite-melteigite series.

Holyokeite—an albitite or albite dolerite with ophitic texture (Holyoke, Massachusetts, U.S.A.).

Invernite—a holocrystalline intrusive rock with phenocrysts of orthoclase and plagioclase in a groundmass of euhedral feldspar and rare hornblende or mica with interstitial quartz.

Josefite—an altered hypabyssal rock composed mainly of augite and olivine with serpentine and calcite.

Kajanite—an effusive rock composed of phenocrysts of bronze-colored mica and olivine in a groundmass of leucite, diopside, and opaque ore; it is feldspar-free.

Kalmafite—a volcanic rock composed of kalsilite and mafic minerals (e.g., katungite, mafurite).

Kamperite—a hypabyssal rock composed of phenocrysts of orthoclase and some oligoclase in a groundmass of dark-colored mica.

Karite—a grorudite with ca. 50% quartz (River Kara, U.S.S.R.).

Kåsenite—a nepheline-bearing carbonatite, rich in pyroxene.

Kassaite—a hypabyssal rock with phenocrysts of haüyne, labradorite, barkevikite, and augite in a tinguaitic groundmass of acicular hastingsite with feldspar.

Katzenbuckelite—a porphyritic rock with phenocrysts of nepheline, biotite, olivine, nosean, leucite, and apatite in a fine-grained groundmass of nepheline, leucite, and acmite with tinguaitic texture (Katzenbuckel, W Germany).

Kirunavaarite—a magnetitite (Kirunavaara, Sweden).

Kodurite—a coarsely crystalline rock of possible (hybrid) igneous origin composed of K-feldspar, garnet, and apatite; Mn-silicate rocks commonly much altered to produce irregularly-shaped, high-grade, Mn-ore bodies (Kodar Mines, India).

Kosenite—see *Kåsenite*.

Kulaite—an effusive rock (nepheline trachybasalt or nepheline basanite) containing orthoclase and calcic plagioclase, small proportions of nepheline and olivine, and hornblende as the dominant mafic mineral (Kula Devit, Turkey).

Kullaite—a porphyritic diorite with altered plagioclase and alkali feldspar phenocrysts in a groundmass with ophitic texture (Kullen, Sweden).

Kuskite—originally a scapolite-bearing porphyry, but the scapolite identified subsequently as quartz (Kuskokwim River, Alaska, U.S.A.).

Kyschtymite—a hypabyssal rock composed of euhedral corundum and some biotite in a groundmass of calcic plagioclase (Kyschtym, U.S.S.R.).

Labrodorite—a leucobasalt; term not in use because of confusion with the mineral (plagioclase) of the same name.

Lehmanite—an igneous rock containing feldspar and quartz.

Leucophyre—(1) altered dolerite with the feldspars saussuritized; (2) a light-colored hypabyssal rock; the antithesis of lamprophyre.

Leucophyride—a field term for a light-colored porphyritic rock with a fine-grained groundmass.

Lindinosite—composed of >50% riebeckite with quartz and microcline.

Lindöite—a variety of sölvbergite characterized by a bostonitic texture (Lindö Island, Norway).

Lundyite—an intrusive rock with orthophyric texture, a high proportion of alkali minerals, and kataphorite-like amphibole (Lundy Island, England).

Magnetitite—an igneous rock composed mainly of magnetite; apatite may be present.

Manjakite—a rock containing garnet, biotite and pyroxene with variable proportions of feldspar and magnetite, and showing a texture resembling that of kentallenite.

Martinite—an effusive rock with vesicular texture, phenocrysts of leucite, feldspar, and augite, and a felty groundmass of labradorite, orthoclase, augite, leucite, olivine, magnetite, and apatite.

Masafuerite—a dark-colored hypabyssal rock composed of abundant phenocrysts of olivine in a groundmass of augite, calcic plagioclase, ilmenite, and magnetite.

Melaphyre—originally applied to any dark-colored porphyritic rock, but later restricted to altered basalt or andesite, particularly amygdaloidal varieties.

Melilitolite—intrusive equivalent of melilitite.

Melmafite—a volcanic rock composed of melilite and mafic minerals (melilitite).

Meymechite (*meimechite*)—an ultramafic rock composed of abundant olivine phenocrysts (usually altered) in a serpentine rock groundmass; considered by petrologists in the U.S.S.R. to be the equivalent of kimberlite.

Microtinite—a coarse-grained leucocratic rock with monzonitic texture and vitreous plagioclase.

Modumite—an anorthosite with plagioclase of bytownitic composition, and some pyroxene and barkevikite.

Nelsonite—refers to a group of hypabyssal dyke rocks composed mainly of ilmenite and apatite, and generally containing rutile (Nelson County, Virginia, U.S.A.).

Nemafite—a volcanic rock composed of nepheline and mafic minerals (nephelinite).

Novaculite—an aphanitic granulose or crypto-crystalline rock mainly composed of quartz, generally with accessory feldspar and garnet.

Orthoclasite—a porphyritic hypabyssal or effusive rock compositionally equivalent to granite or syenite in which orthoclase is commonly > 90%.

Orthofelsite (*orthophyre*)—a porphyritic rock with phenocrysts of orthoclase, but not quartz, in a felsitic groundmass.

Orthosite—a light-colored, coarse-grained rock composed almost entirely of orthoclase.

Orvietite—an effusive rock composed of plagioclase and sanidine in approximately equal proportions, with leucite, augite, and minor biotite and olivine.

Ostraitite—a jacupirangite with abundant green spinel (Ostraia Sopka, U.S.S.R.).

Pawdite—a melanocratic hypabyssal rock mainly composed of magnetite, sphene, biotite, hornblende, and calcic plagioclase with bytown-

ite cores and rimmed by orthoclase.

Penikkavaarite—an intrusive rock composed of augite, barkevikite, and green hornblende in a feldspathic groundmass.

Plagiophyre—a porphyritic rock resembling orthophyre but with plagioclase rather than orthoclase phenocrysts.

Plumasite—a hypabyssal rock of variable composition with allotriomorphic-granular texture and mainly composed of corundum crystals enclosed in oligoclase.

Pollenite—compositionally similar to tautirite but containing olivine and having a glassy groundmass.

Protectite—a rock formed by crystallization of a primary magma.

Puglianite—a coarse-grained rock composed of augite, leucite, anorthite, sanidine, hornblende, and biotite.

Queluzite—mainly composed of spessartine with some amphibole, pyroxene, or mica.

Quenite—a dark-colored hypabyssal rock composed of anorthite and Cr-diopside with smaller proportions of olivine and bronzite.

Redwitzite—a melanocratic biotite-bearing monzonite; a plutonic equivalent of kersantite; found in the Hercynides of Europe, possibly as part of the durbachite suite (see **Durbachite**).

Riedenite—large tabular biotite crystals in a groundmass of nosean, biotite, pyroxene, and small proportions of sphene (titanite) and apatite.

Rougemontite—coarse-grained and mainly composed of anorthite and titanaugite; small proportions of olivine and opaque ores also present (Rougemont, Montreal, Canada).

Routivarite—a fine-grained rock composed of orthoclase, plagioclase, quartz, and garnet (Routivara, Sweden).

Sanidinitite—a name variously used for rocks mainly composed of sanidine or other forms of alkali feldspar.

Sannaite—an effusive rock with phenocrysts of barkevikite, pyroxene, and biotite in a fine-grained groundmass of alkali feldspar, acmite, chlorite, calcite, and pseudomorphs of mica after nepheline.

Schorenbergite—a tinguaitic hypabyssal rock containing phenocrysts of nosean or leucite in a groundmass of leucite, nepheline, and acmite.

Sebastianite—a plutonic rock composed of anorthite, biotite, some augite, and apatite; neither feldspathoids nor quartz are present.

Selbergite—a hypabyssal rock with abundant phenocrysts of leucite, nosean, sanidine, acmite-augite, and biotite in a groundmass of nepheline, alkali feldspar and acmite (Selberg, W Germany).

Skomerite—an effusive rock containing phenocrysts of augite, olivine, and altered plagioclase

in a plagioclase-rich groundmass (Skomer Island, Wales).

Spinellite—a medium- to coarse-grained titaniferous, magnetite-rich rock with hypidiomorphic-granular texture and spinel up to 20%.

Sumacoite—a magnetite-rich effusive rock with abundant phenocrysts of plagioclase and augite, and rare olivine, in a groundmass of andesine, orthoclase, nepheline, and haüyne (Sumaco crater, Ecuador).

Sussexite—a porphyritic tinguaite rock composed mainly of nepheline and acmite (Sussex County, New Jersey, U.S.A.).

Tamaräite—a dark-colored hypabyssal rock resembling basalt in appearance; it consists of augite, hornblende, biotite, nepheline, plagioclase, orthoclase, and minor accessory minerals with secondary cancrinite and analcime.

Tarantulite—a plutonic rock transitional between alaskite and silexite that contains >50% quartz, orthoclase and albite and up to 5% dark minerals (Tarantula Spring, Nevada, U.S.A.).

Tautirite—a rock composed of K-feldspar, andesine, nepheline, and amphibole with abundant accessory sphene (titanite).

Tavolaitite—a porphyritic rock with large phenocrysts of leucite, haüyne, acmite–augite and garnet in a groundmass of the same minerals with interstitial leucite, orthoclase, haüyne, labradorite, acmite–augite, biotite, garnet, and nepheline (Tavolato, Italy).

Taxite—a volcanic rock that appears to be elastic as the result of the mixture of materials of varying texture and structure from the same flow; where the components are mixed in a breccia-like manner, the term *ataxite* is used; where the components are aggregated into separate bands, the term *eutaxite* is used.

Tollite (toellite)—a hypabyssal rock with phenocrysts of hornblende and andesine and lesser proportions of biotite, white mica, quartz, and feldspar, but with no dark-colored minerals in the groundmass.

Topazite (topazfels, topazogene)—a hypabyssal rock almost entirely composed of quartz and topaz.

Toryhillite—a plutonic rock (albite-rich urtite) composed of albite, nepheline, clinopyroxene, garnet, opaque ores, calcite, and sometimes sodalite (Toryhill, Ontario, Canada).

Trowlesworthite—a coarse-grained plutonic rock with red orthoclase, tourmaline, fluorite, and quartz; probably represents granite affected by pneumatolysis (Trowlesworth, Cornwall, England).

Tveitåsite—a dark-colored rock, found at some igneous contacts, composed mainly of alkali feldspar with accessory sphene and apatite (see **Fenite**).

Ulrichite—a dark-colored hypabyssal rock with alkali feldspar, Na-pyroxene, amphibole, and smaller and less common olivine phenocrysts in a groundmass of feldspar, pyroxene, and amphibole; an olivine-bearing tinguaitite.

Varnsingite—a coarse-grained dyke rock containing >50% albite, pyroxene, sphene, magnetite, apatite, and secondary epidote, prehnite, chlorite, amphibole, and muscovite; possibly a pegmatoid associated with gabbro.

Vaugnérite—a melanocratic diorite containing biotite, hornblende, plagioclase, quartz, and accessory apatite, magnetite, pyrite, and sphene; a plutonic equivalent of minette found in the Hercynides of Europe, possibly as part of the durbachite suite (Vaugnéray, France—see **Durbachite**).

Venanzite—a holocrystalline porphyritic effusive rock with phenocrysts of olivine and phlogopite in an aphanitic groundmass of these minerals with melilite, leucite, and magnetite (San Venanzo, Italy).

Ventrallite—an alkalic rock with more K-feldspar than calcic plagioclase; nepheline is the essential feldspathoid.

Vintlite—a porphyritic hypabyssal rock with phenocrysts of labradorite or bytownite and hornblende in a groundmass of feldspar, hornblende, and quartz; compositionally similar to a hornblende diorite.

Volcanite—(1) a volcanic rock; (2) an effusive rock with phenocrysts of anorthoclase, andesine, and augite in a glassy groundmass containing feldspar and augite microlites.

Wennebergite—a quartz porphyry with phenocrysts of orthoclase, biotite and quartz in a microcrystalline groundmass (Wenneberg, W Germany).

Wernerite—a light-colored rock composed almost entirely of scapolite and recorded as being of igneous derivation.

Wesselite—a hypabyssal rock containing biotite, barkevikite, titanaugite, haüyne, and nepheline (Wesseln, Czechoslovakia).

Westerwaldite—an effusive rock with phenocrysts of serpentinized olivine, some with rims of augite, in a groundmass of labradorite, sanidine, augite, and biotite with interstitial nepheline.

D. R. BOWES

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IGNEOUS ROCKS—TECTONIC SETTING

Igneous rocks are products of partial melting in the upper mantle, subducted oceanic crust, or lower continental crust. Individual igneous suites may be produced exclusively by one such process, or may represent the product of more than one operating in concert. Within a plate tectonic framework two environments for the generation and emplacement of igneous rocks are recognized: at plate margins and within plates. These in turn, can be subdivided into eight tectonic settings (Fig. 1).

Plate margins

- Convergent(destructive) 1. Ocean–ocean island arc
2. Continent–ocean arc
Divergent(constructive) 3. Mid-ocean ridge
4. Back-arc basin

Within plate

- Tensional 5. Intracontinental rift
6. Intramontane rift

Neutral

7. Oceanic hot spot (ocean island)
8. Continental hot spot

Convergent margins

Convergent margins are the sites where oceanic crust returns to the upper mantle by subduction, beneath the leading edge of either an oceanic or a continental plate. Magma may be generated by partial melting of the down-going slab, the overlying mantle wedge, and the basal continental crust. The contribution of each possible source influences the composition of the resulting igneous rocks. Two major variants are recognized, both dominated by tholeiitic and calc-alkaline rocks (Fig. 2; see **Circum-Pacific plutonic and volcanic activity**).

1. *Ocean–ocean collision zones* occur where ocean crust is subducted beneath ocean crust, e.g., the island arcs of the western Pacific Ocean from the Kermadec arc (N of New Zealand) to the Bonin–Izu arc (S of Japan). Others include the western Aleutian arc, the Andaman arc, the Scotia arc, and Lesser Antilles arc (Fig. 3). Both tholeiitic and calc-alkaline rocks are present, but the proportions differ when compared to the continent–ocean arcs; basalts and andesites predominate, while dacites and rhyolites are relatively rare. In some cases even andesites are uncommon. Boninite, a Mg-rich mafic andesite, is unique to these arcs. Plutonic rocks are not abundant; gabbro and diorite predominate, with subordinate plagiogranites.

These are the true, festoon island arcs, with

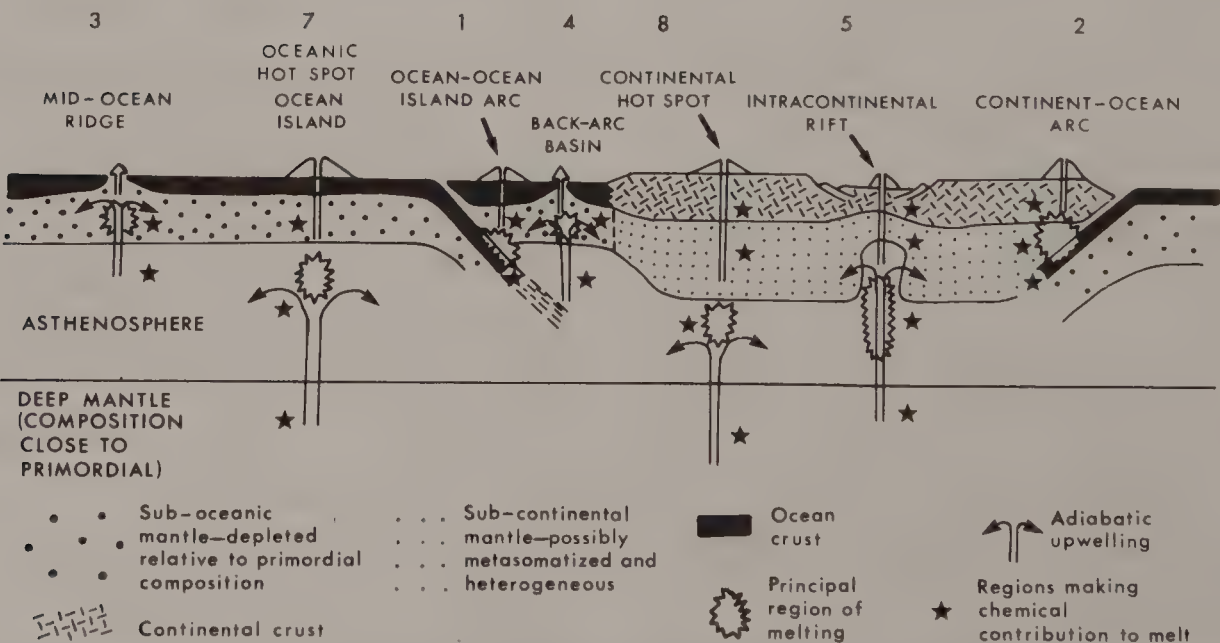


FIGURE 1. Synoptic diagram illustrating the role of partial melting of various source rocks and sources of contaminants during magma generation in igneous rocks in the principal plate tectonic settings; 1–5, 7, 8 refer to tectonic settings set out in the text.

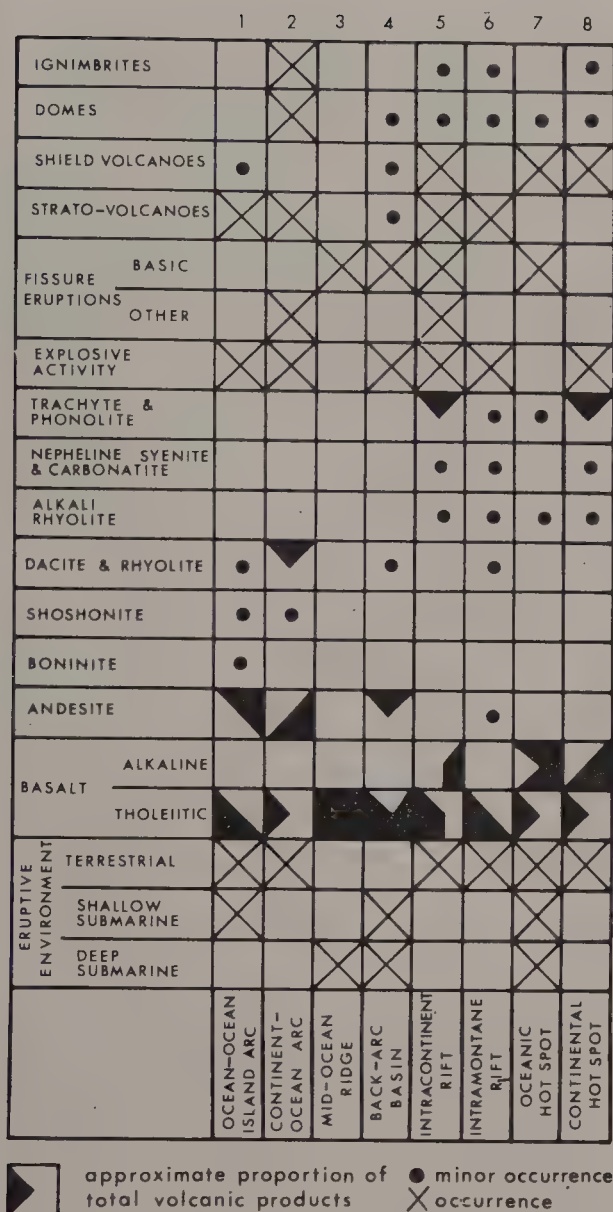


FIGURE 2. Tabulated characteristics of igneous provinces in the eight principal plate tectonic settings.

only the tops of the largest volcanic edifices appearing above sea level. This bathymetry reflects an isostatic equilibrium, indicated by the abundance of hyaloclastic rocks throughout the volcanic edifice, suggesting that a large percentage of the history of eruption occurred close to mean sea level. Explosive volcanic activity only accompanies the later andesitic or more acidic phases of activity.

2. *Continent-ocean collision zones*, where oceanic crust is subducted beneath continental crust, are typified by the Andean belt from Colombia to Tierra del Fuego. Other examples include the Sunda arc in Indonesia, the Japan and Kurile-Kamchatka arcs, the eastern portion of the Aleutian arc and the Cascade range, all around the N Pacific, and the New Zealand arc

(Fig. 3). These include some of the most volcanically and siesmically active areas on Earth.

The complete suite of tholeiitic to calc-alkaline igneous rocks is represented by basalt, andesite, dacite, and rhyolite amongst the lavas and gabbro, diorite, granodiorite, tonalite, and granite among the intrusions. Andesite and dacite dominate the lavas, while the plutonic rocks are dominated by granodiorite and tonalite. The geological record indicates a preponderance of terrestrial over submarine eruptive rocks, and the lavas are interlayered with pyroclastic rocks in large stratovolcanoes. Comagmatic plutons and lavas have been recognized in western N America and the Andes, where plutonic rocks often intrude the volcanic pile. Cauldron subsidence accompanied by very violent explosive volcanic activity is characteristic, and the products of plinian and peléan eruptions, like ignimbrites, are widespread (see **Explosive volcanic eruptions—classification**).

Both types (1 and 2) of convergent margin possess a fundamental asymmetry, reflecting the presence of the seismic Benioff zone dipping away from the trench, marking the downward path of the subducted slab. The existence of a systematic variation in composition perpendicular to the length of the igneous arc has proved controversial over the last 20 years (see review in Miyashiro et al., 1982).

Divergent margins

Divergent margins are the sites of ocean-floor spreading located at the mid-ocean ridges and the spreading axes in some back-arc basins (e.g., the Scotia Sea and Sea of Japan). The magma-generating process is located in the upper mantle and produces a tholeiitic basalt liquid. The subsequent differentiation of this produces the observed igneous suite, while the mechanism of ridge spreading is considered to reflect the structural relationship and emplacement modes observed in ophiolites (obducted remnants of oceanic or back-arc basin crust).

3. *Mid-ocean ridges* are dominated by tholeiitic basalts erupted in deep water as both pillow lavas and massive flows. There exists a spatial variation, often between different segments of the same ridge, between high-Mg and low-Mg basalts that may reflect the rate of sea-floor spreading. Deep-sea drilling has proved the existence of dolerite dykes beneath the lava layer, as suggested by the structure of ophiolites. Off-axis volcanic activity, whose relationship to the magmatism at the axis is obscure, produces a high-Mg basalt suite that may include basaltic komatiites and picrites.



FIGURE 3. Distribution of active and recently active volcanic provinces with the present plate configuration.

4. *Back-arc basin spreading axes* are the sites of similar magmatism to that seen in the oceans, although variations include the presence of a high-Al group of basalts produced in the early stages of back-arc rifting. Eruptions can occur initially in shallow water and, as a consequence, hyaloclastic rocks may be interbedded with pillow lavas and massive flows.

The nature of the lower layers of newly generated crust in the ocean and marginal seas can only be deduced from studies in ophiolites and the seismic properties determined geophysically. As such, the igneous rocks include those forming tectonized mantle (lherzolite), the residual material from partial melting (harzburgite), the axial magma chamber (massive and layered gabbros with their ultramafic cumulates), and its differentiates (diorite and plagiogranites). Much of the ultramafic material is serpentinized and the dykes and lavas are spilitized or show the effects of ocean-floor metamorphism (see **Ophiolite; Ocean-floor metamorphism**).

Tensional within-plate provinces

Two quite distinct provinces can be defined, unified simply by their association with rift systems.

5. *Intracontinental rift systems*, such as those

of E Africa and the Rhine Valley in W Europe, are characterized by igneous rocks showing a great range and often extreme compositions. Other examples include the Baikal rift in Siberia and the older Keweenaw "rift" in N America. The bulk of the igneous rocks form lava plateaus, largely composed of both tholeiitic and alkaline basalts. The early activity is related to fissures, but later evolution of a province sees activity concentrated in discrete centers, often shield volcanoes. Such centers are invariably complex, often with subvolcanic ring complexes composed of strongly differentiated igneous rocks. Nepheline-normative basalts give way to phonolites and trachytes, or even peralkaline foidal varieties of bewildering diversity. The most spectacular of these are the extremely sodic and extremely potassic rocks, often rich in carbonate. The carbonatite complexes of E Africa are considered to be the subvolcanic equivalents of such unusual volcanoes as Oldoinyo Lengai (Tanzania), which erupt sodium carbonate lavas and ash.

The early stages of continental break-up involve the development of intracontinental rift systems. These evolve from a series of "triple junctions" into a fully fledged spreading ocean basin with the third member of each triplet constituting a "failed arm." The evolution of

igneous provinces in a triple-armed rift system in which two arms open into an ocean basin and one "fails" is quite different. All three arms share the early tholeiitic basaltic phase, but usually only the failed arm develops as an alkaline or peralkaline province. Such a relationship is well-illustrated in the early Tertiary province of E Greenland, where the tholeiitic to alkaline basalts and layered intrusions (e.g., Skaergaard) follow the two arms of a triplet that opened into the N Atlantic Ocean, but the failed inland arm contains a series of syenite and nepheline syenite intrusions.

6. *Intramontane rift systems* occur in the tensional environment of the uplift plateaus of Southwestern U.S.A. (Basin and Range province and the Rio Grande rift) and Tibet. Again, strongly differentiated igneous rocks are abundant, but the typical suites are bimodal, between tholeiitic or calc-alkaline basalts and andesites and acidic, often sodic rocks, generally rhyolites and obsidian. The eruptive activity, though not intense, may be violently explosive with ignimbrite and collapsed cauldron structures being common, e.g., Valles Caldera, New Mexico.

Hot spots

Upwelling, deep-seated mantle plumes manifest themselves as aseismic areas of high heat flow and volcanic activity within both continental and oceanic plates. In the volcanic areas they are marked by linear volcanic provinces representing a migrating volcanic center on a lithospheric plate moving over a fixed mantle plume (e.g., the Hawaiian chain and the Emperor seamounts). On continental areas, hot spots have been invoked to explain the domed areas of uplift associated with triple junctions in rift systems, and such cases are indistinguishable from rift valley igneous provinces. However, there exist igneous provinces within continental areas (often cratonic) where igneous activity does not appear to be related to rift systems.

7. *Oceanic hot spots* are best documented in the case of the Hawaiian Islands. The volcanism produces immense shield volcanoes composed largely of alkaline basalts, which are often nepheline-normative. These volcanic edifices are built up from the abyssal plain, so that pyroclastic rocks only occur towards the top of the whole pile where eruptions occurred close to sea-level. The late stages in the evolution of these shield volcanoes sees the eruption of more strongly differentiated magmas, with trachytic rocks (mugearite, hawaiite, and trachyte itself) being commonly developed as domes and parasitic cones.

Somewhat more complex relationships are seen on the oceanic islands of the Atlantic Ocean, especially those related to hot spots along and offset from the mid-Atlantic ridge (Iceland, Azores, and Tristan da Cunha). Basalts predominate, but the volcanic piles are often capped by an alkaline or subalkaline suite of lavas. In a few cases there are intrusions of sodic granite. Three suites are recognized: (1) tholeiitic basalt followed by rhyolite and comendite with few intermediate rocks; (2) alkaline basalts followed by basanite, tephrite, and pantellerite; and (3) alkaline basalt followed by hawaiite, mugearite, and trachyte or phonolite.

8. *Continental hot spots* are associated with lava plateaus and both tholeiitic and alkaline basalts, predominantly the latter, occur. As the province evolves, more strongly differentiated magmas are evident, with peralkaline or strongly acidic lavas being erupted, e.g., trachyte and phonolite, or rhyolite and comendite. Subvolcanic ring complexes often include rocks of extreme compositions, an extreme example being the Devonian ring complexes with their satellitic nepheline syenites and carbonatites in the Kola peninsula, U.S.S.R. Where these differentiated rocks occur, extremely violent explosive volcanic activity may be associated with caldera collapse, e.g., Yellowstone area, Wyoming, U.S.A.

The Canary Islands present an intriguing case for the influence of oceanic or continental basement on igneous rocks associated with a hot spot. All the islands are basaltic shield volcanoes capped by alkaline lavas, but Fuerteventura and Lanzarote are on continental basement and include a more strongly differentiated alkaline rock suite. A similar petrographic duality is seen in the igneous rocks of the Cameroon line, where one hot spot has migrated beneath both continental and oceanic crust.

Cratonic igneous rocks

Within the continental cratons there exist both discrete igneous provinces and scattered occurrences of particular igneous rocks whose tectonic setting is obscure. Some show a restriction to the older cratons but others are more widespread. The most conspicuous are the kimberlites and associated rocks of southern Africa and the formerly adjacent areas of Gondwanaland. These include kimberlite, lamprophyres and lamproites, melilitic rocks, and carbonatites. As a suite they are restricted to the Precambrian shield, and within this suite a diamondiferous group is restricted to the Archaean areas of the shield. Other features of

these stable shield areas are the large stratiform layered basic intrusions, of which some (e.g., Muskox intrusion, N Canada, Nizhni Taghil in Siberia, U.S.S.R., and the Great Dyke of Zimbabwe) may have been emplaced in deep continental rift zones related to flood basalt eruptions; others, like the Bushveld Complex, S Africa, have a more obscure tectonic setting.

Geochemical trends

Hatherton and Dickinson (1969) were the first to show that K₂O content at given SiO₂ levels in modern calc-alkaline volcanic rocks increases with depth to the descending subducted ocean plate. Other large-ion lithophile (LIL) elements in these rocks show similar trends. Rare-earth element (REE) patterns also change from flat to light REE-enriched, the K/Rb ratio decreases and the range of ⁸⁷Sr/⁸⁶Sr ratios increases with this increasing depth. This compositional polarity in island and cordilleran arcs is related to one or more factors (Condie, 1982): (1)

progressive changes in the dominant solidus phase in magma source areas—changing from amphibole to clinopyroxene as the subducted slab descends; (2) small degrees of melting at greater depths in a progressively dehydrating descending slab; (3) greater contamination of magma ascending through more overlying mantle and/or continental crust as the slab descends; or (4) larger amounts of K₂O leached by water as this is driven off the descending slab and interacts with the overlying mantle wedge.

The degree of differentiation in such suites may reflect the age of the subducted crust and the speed of subduction (Fig. 4). Trace-element and Sr, Pb, and Nd isotope data favor an origin for subduction zone-related, calc-alkaline magmas by partial melting of mafic rocks in the descending slab with varying contributions from the overlying mantle wedge and any older crust above.

Mid-ocean ridge basalts (MORB) and those from back-arc (marginal) basin spreading centers are chemically very similar. They have a

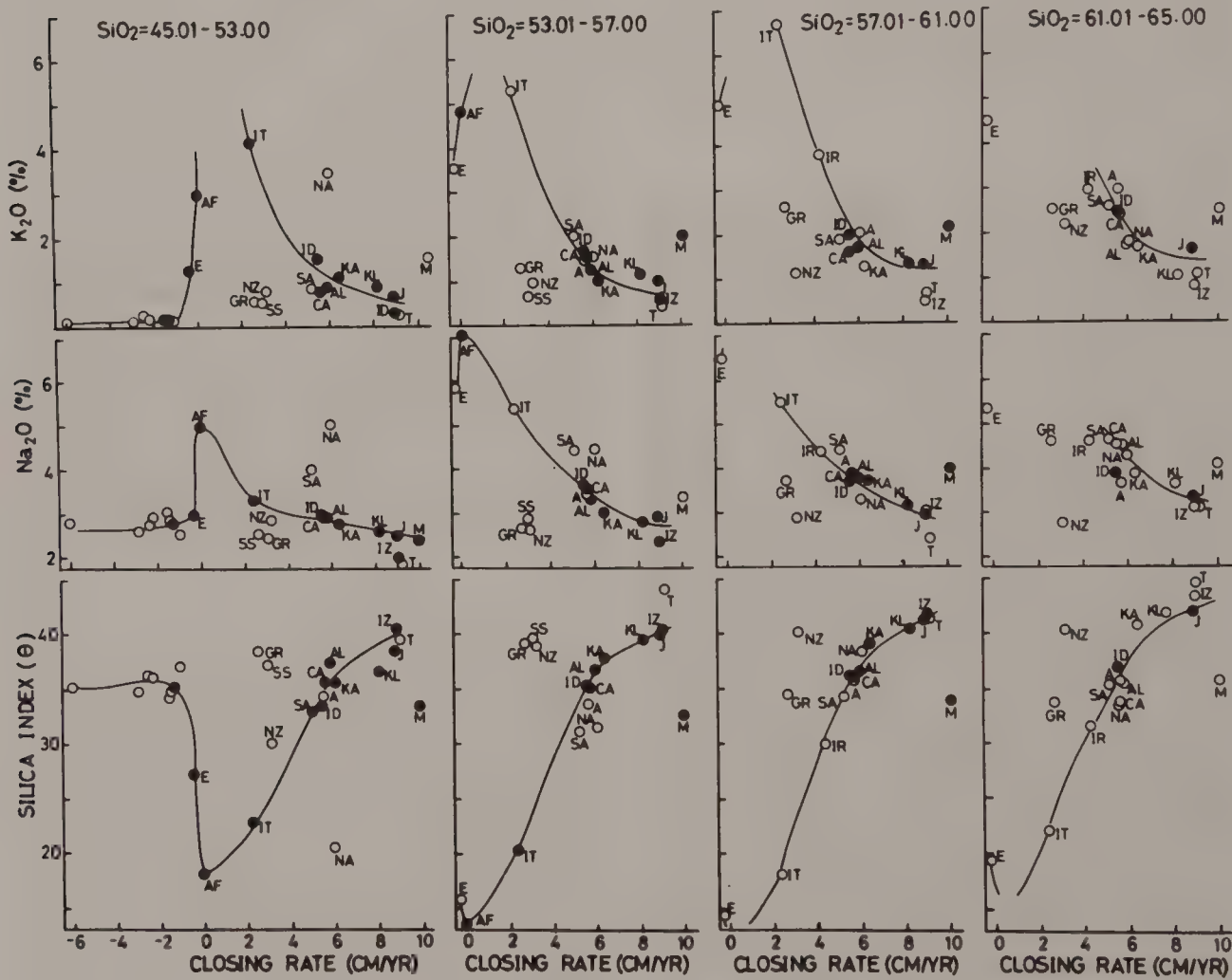


FIGURE 4. Relationship of chemical composition of young volcanic rocks to rates of plate motion (+, consumption; -, creation): ●, oceanic volcanic rocks; ○, volcanic rocks from continent-ocean margins. (From Sugisaki, 1976.)

lower LIL content than tholeiitic basalts from other settings, flat REE patterns (ca. $10 \times$ chondrite) and low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. This, together with evidence from experimental studies, implies an origin in the upper mantle involving 20–30% partial melting of a lherzolite, itself depleted with respect to primitive mantle chondrite. After melting, the predicted olivine + pyroxene residue approximates to harzburgite. Further melting of this residue may account for the high-Mg picritic and komatiitic rocks sometimes seen in ophiolites.

Ocean-island basalts (OIB), when compared to MORB, are enriched in LIL elements, exhibit slightly light REE-enriched patterns and have higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. These data suggest partial melting of variable amounts of a lherzolitic source closer to the primitive chondrite composition, such as may be brought up by hot plumes from the deep mantle (Sun and Hanson, 1975).

Basalts associated with intracontinental rifts are highly enriched in LIL elements and light REE elements compared to both MORB and OIB. They also have a large range in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Furthermore, they show a considerable range, both between provinces and within single rift systems. Multistage models seem necessary to explain all the petrogenetic diversity, involving varying degrees of partial melting of mantle peridotites, fractional crystallization of the resultant magma, crustal contamination and wall-rock reaction (Condie, 1982). This diversity in rift-valley rocks may also be influenced by gross inhomogeneity in the subcontinental mantle (Fig. 1).

The strongly bimodal rocks of intramontane rift systems on the great plateau areas reflect magmas enriched in LIL and light REE elements. They exhibit highly variable $^{87}\text{Sr}/^{86}\text{Sr}$ ratios with the extremely felsic varieties having the highest values; the range also shows some degree of bimodality. This is considered to reflect a mantle origin for the basic members, and a lower crustal melt origin for the felsic varieties.

Geochemical discrimination for tectonic setting

Tholeiitic basalts can occur in most of the aforementioned tectonic settings, and since 1970 considerable effort has been devoted to defining geochemical parameters for their discrimination. Early attempts at this were largely empirical, defined using a trace-element data set gathered from basaltic lavas erupted in known tectonic settings—supposedly “immobile” trace elements, i.e., Ti, Zr, Sr, Y, and Nb, were used (Fig. 5; cf., Pearce and Cann, 1973; Winchester and Floyd, 1976). Some statistical refinement of

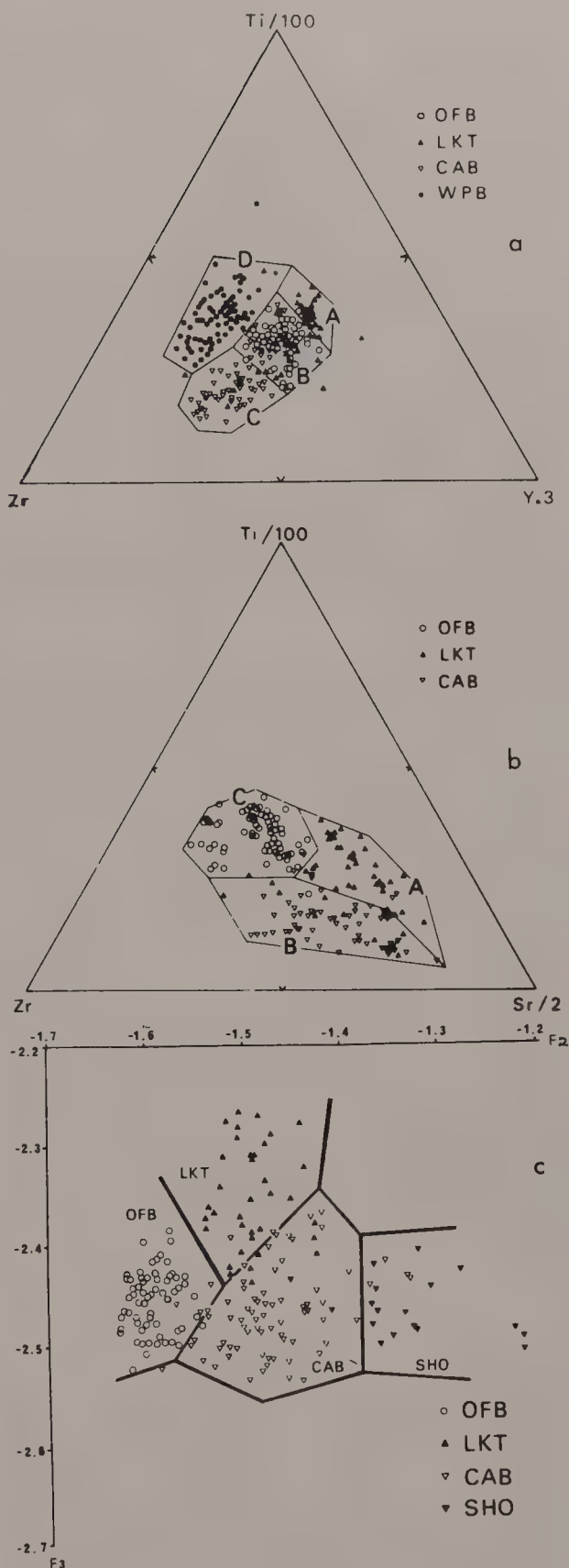


FIGURE 5. Trace and major element discriminant diagrams for basic volcanic rocks: (a) Ti–Zr–Y, (b) Ti–Zr–Sr, (c) major element discriminant function. OFB, ocean-floor basalt; LKT, low-K tholeiite; CAB, calc-alkaline basalt; WPB, within-plate basalt; SHO, shoshonite. ((a) and (b) from Pearce and Cann (1973), (c) from Pearce (1976).)

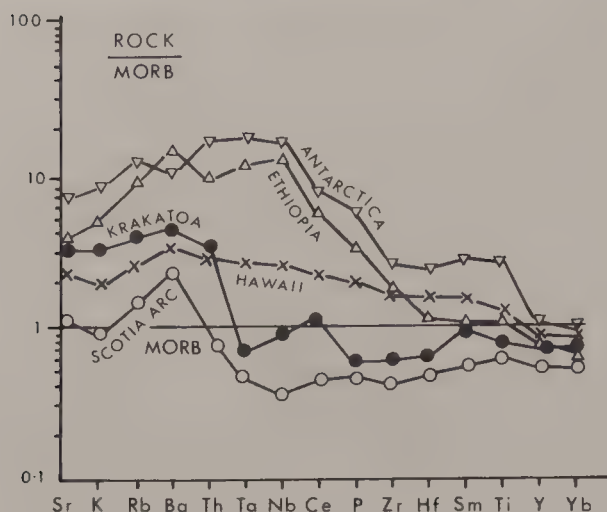


FIGURE 6. MORB-normalized incompatible element spidergram illustrating differences between basalts in five tectonic settings compared to those erupted at mid-ocean ridges (MORB). (After Pearce, 1983.)

this approach was achieved using discriminant function analysis (Pearce, 1976).

A more sophisticated approach seeks to define a petrogenetic basis for these empirical patterns, and this has concentrated on combining a knowledge of the behavior of incompatible trace elements in the supposed melt-generating process with magmatic modeling techniques. This not only permits a trace element characterization of basalts produced by subduction and the partial melting of enriched and depleted mantle, but also allows more subtle interpretation of petrogenesis in various basalt provinces, recognizing such features as crustal contamination and mantle heterogeneity (Fig. 6). Discriminant criteria are a by-product of this approach (cf., Pearce and Norry, 1979; Pearce, 1983), which has also demonstrated the need to use trace-element and isotope geochemistry in concert.

K. C. CONDIE

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Cross-references: *Andesite line; Archaean volcanism; Basalt; Calc-alkaline rocks; Carbonatite; Circum-Pacific plutonic and volcanic activity; Flood basalt; Island arcs—petrology; Kimberlite; Mid-ocean ridge and ocean-floor petrology; Ocean-island igneous rocks; Ophiolite; Pacific and Atlantic suites; Peralkaline igneous rocks; Rift valley volcanism; Steinmann Trinity; Ultramafic lavas.*

INJECTION COMPLEX

An *injection complex* is an assemblage or association of rocks consisting of granitic material in intricate relationship with metamorphic or sedimentary rocks. The term is widely used for those migmatite complexes, or parts of such complexes, in which neosome material is both abundant and vein-like (Fig. 1). It is particularly used as a descriptive field term without genetic connotations concerning neosome derivation, although in many cases the “injected” material is interpreted as being of igneous derivation, particularly residual fluids from a granitoid mass. This was how Read et al. (1925) and Read (1931) interpreted the injection complexes of Sutherland in northern Scotland when the term was introduced. One of the injection complexes there was divided into a zone of veins, a zone of injection and a central granitic mass (Fig. 2). The zone of injection consists of migmatites and all “of Sederholm’s rock varieties, arterites, eruptive-breccias (agmatites), banded gneisses, pygmatitic migmatites, diktyonites, nebulites and the like can be paralleled” (Read et al., 1925). Especially referred to are the processes of permeation and injection. In the former, granitic or pegmatitic material is considered to soak into the country rocks to produce banded granites in which the structures of the preexisting rock remain as “replacement” structures. Injection was envisaged as the intrusion of “granitic” liquid of such extreme fluidity that it was able to both

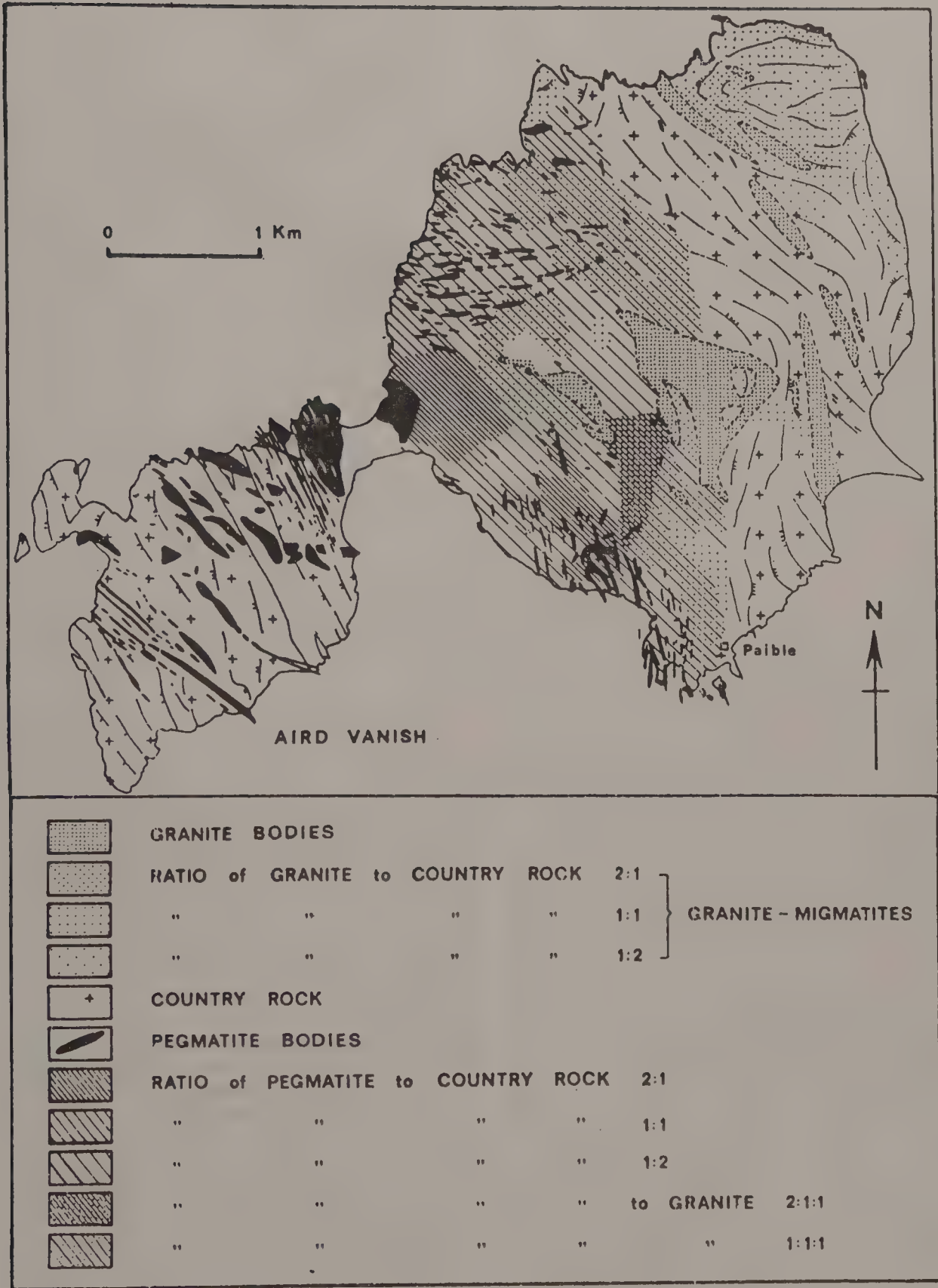


FIGURE 1. Injection complex, Taransay, Outer Hebrides, Scotland. (After Myers, 1971.)

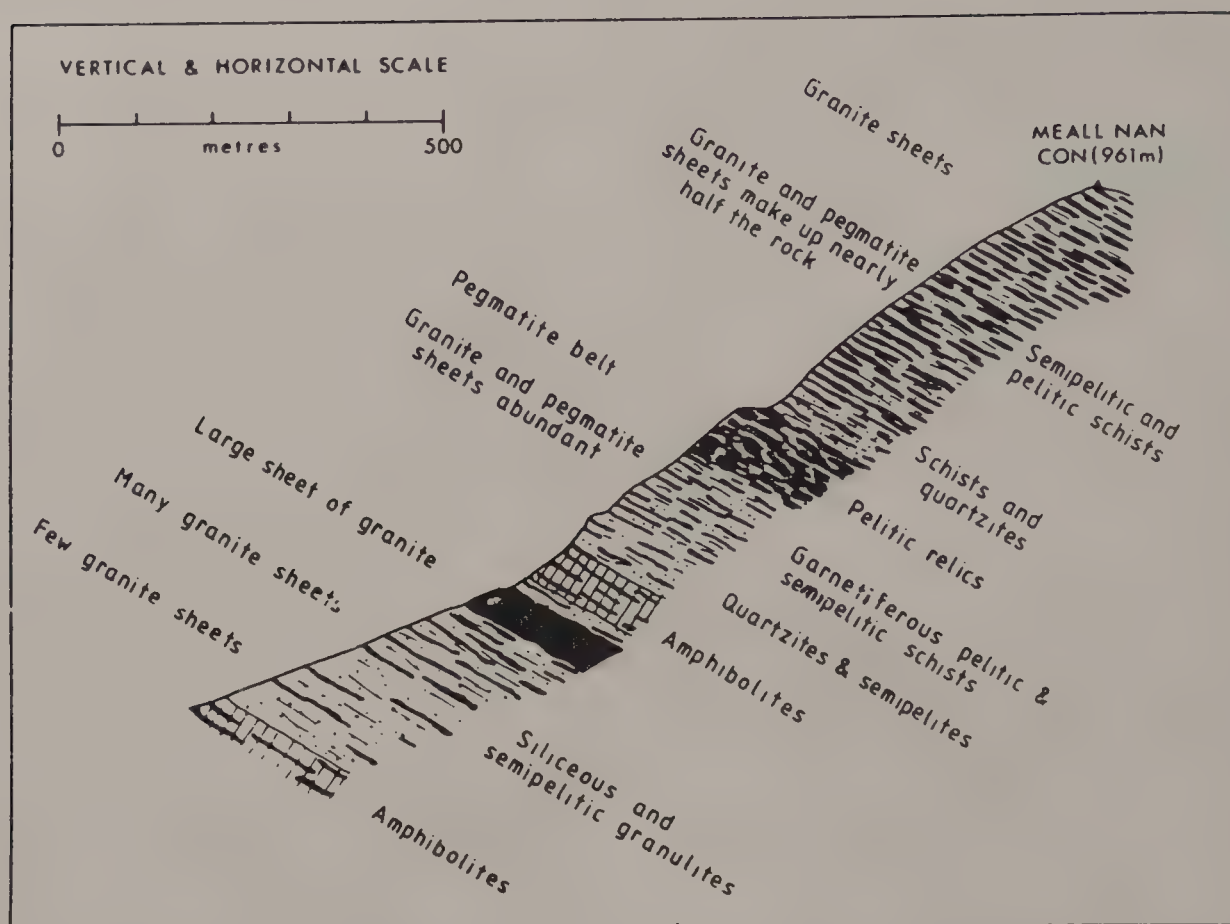


FIGURE 2. Cross-section of injection complex, Ben Klibreck, Sutherland, Scotland. (From Hall, 1987.)

penetrate the country rocks in "lit-par-lit" manner and to produce many sorts of veins (see **Gneiss**).

Read, who was familiar with the ideas of A. Michel-Lévy, wherein the ascent of granitic magma is a kind of fusion preceded by the action of heat and mineralizers, as well as with the ideas of palingenesis expounded by J. J. Sederholm, nevertheless proposed a somewhat conservative explanation of the Sutherland phenomena—the migmatites were formed by the action of residual fluids from the crystallizing central granodiorite that had been intruded into country rocks during the last stages of metamorphism. Subsequent workers have suggested different processes; Cheng (1943) envisaged the influx of a succession of alkalic and other fluids that accomplished a series of replacements and produced migmatitic rocks, and Ma (1948) presented evidence to support the view that the agmatites of the complex were formed in situ as a result of recrystallization dependent on migrating chemical elements "probably in the ionic state."

Local partial melting is now the favored method for neosome development in injection complexes in migmatite complexes, with Camp-

bell (1980) demonstrating the successive development of neosomes in a metamorphic regime that reached P - T conditions of 825°C - 6.5 kb . This stage by stage development of the neosome material amongst polyphase metamorphosed and polyphase deformed rocks (Hopgood, 1984) making up injection complexes has been confirmed by isotopic studies (e.g., Hopgood et al., 1983). However, in the literature many migmatite complexes have been treated as if there was a single stage of injection into metamorphic rocks affected by a single stage of recrystallization, and involved arguments have been presented for the apparently "anomalous" mineral associations.

Locally structurally-controlled emplacement at the margin of a pluton, with the development of parallel sheets or fingers of granitic material in the rocks of the contact aureole, results in the development of what can be termed an injection complex, but is better referred to as an interfingering igneous contact zone.

D. R. BOWES

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Cross-references: *Anatexis*; *Emplacement—modes*; *Gneiss*; *Granite*; *Metasomatism*; *Migmatite*; *Migmatite—origin of neosome*; *Migmatite—structural relations*; *Pegmatite*.

INTERMEDIATE IGNEOUS ROCKS

The general and widely used term *intermediate igneous rocks* applies to igneous rocks with SiO₂ contents intermediate between those of acidic and basic igneous rocks. The limits of 52–65 wt.% SiO₂ set by Rosenbusch (1898) are widely accepted, although this range excludes a few extreme members of certain igneous families that otherwise lie within the intermediate range. Petrographers apply the term loosely and directly to members of the families given in Table 1, often on purely petrographic grounds without knowledge of the SiO₂ content from analysis. Other general terms in igneous rock classification cut across and are

complementary to the acid–intermediate–basic grouping. Intermediate rocks may be oversaturated (but never strongly so) or undersaturated (sometimes extremely so) and while the former may be intermediate between acidic and basic in terms of dark mineral content, the latter may be exceptionally leucocratic rocks. For these reasons, and because of the genetic connotation of “intermediate,” the term tends, in modern literature, to be most commonly applied to the dioritic–andesitic and monozonitic intermediate rocks (which can be genetically intermediate between basic rocks and acidic derivatives), while saturated and undersaturated syenitic rocks (which are generally residual rocks, in a genetic sense) tend to be set apart by terms such as “alkaline” or, where appropriate, “feldspathoidal,” although they strictly lie within the specified intermediate SiO₂ limits (see **Alkaline rocks—undersaturated**; **Feldspathoidal rocks—genesis**).

The shortcomings of the term when used in a precise sense are discussed by Hatch et al. (1956). Johannsen (1937) confuses the issue by including gabbros in his volume III on “Intermediate Rocks” while quoting the Rosenbusch limits in his glossary, volume I. The term does not appear in the Streckeisen (1976) classification and some or all members of fields 4–15 of this classification will be intermediate rocks. The most important types are listed in Table 1, with the nature of the feldspar as the distinction between alkaline and calc-alkaline types, the most important subdivision within the intermediate range. A profusion of special named varieties of these named intermediate rocks exists, particularly of the alkaline members; this stems in part from the nature of feldspar variation in these rocks (see, for example, **Syenite**). The granodiorites straddle the upper limits of the intermediate range, and in broad descriptive work are often referred to as acidic rocks.

Intermediate rocks are probably relatively less abundant in the crust than acidic or basic rocks (see Richardson and Sneesby, 1922) although the magnitude of this gap is difficult to quantify and may well have been exaggerated.

TABLE 1. Main types of intermediate igneous rocks

	Alkaline		Calc-alkaline	
	Feldspathoid-bearing	Alkali feldspar >65%	Alkali feldspar and plagioclase 35–65%	Plagioclase feldspar >65%
<i>Plutonic</i>	Nepheline syenite	Syenite	Monzonite (Syenodiorite)	Diorite (Granodiorite)
<i>Extrusive</i>	Phonolite	Trachyte	Trachyandesite (Latite)	Andesite (Dacite)

The calc-alkaline types are by far the most abundant. In tectonically active regions rocks of the quartz diorite–granodiorite series make up a major part of the huge composite “granite” batholiths. Truly acidic members of such bodies are comparatively uncommon.

Great thicknesses of andesitic extrusive rocks also characterize orogenic environments; these may grade toward the basic end of the intermediate range, into tholeiitic basalts. Alkaline intermediate rocks are volumetrically rare rocks, although of great mineralogical diversity and hence complexity in nomenclature. As shallow plutonic bodies they are characteristically found in regions of large-scale continental rifting. Common associations are minor bodies of alkali gabbro, carbonatites, and peralkaline acid rocks. Alkaline intermediate derivatives sometimes occur as minor members of dominantly gabbroic intrusive bodies. The extrusive equivalents (trachytes, phonolites) are also most prominent in stable continental regions, yet also occur in association with basalts in many mid-oceanic islands.

IAN PARSONS

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Cross-references: *Acid(ic) igneous rocks*; *Alkaline rocks—undersaturated*; *Andesite and dacite*; *Basic igneous rocks*; *Calc-alkaline rocks*; *Feldspathoidal rocks—genesis*; *Latite*; *Monzonite (syenodiorite)*; *Plutonic rocks*; *Syenite*; *Trachyandesite*, *Trachybasalt*; *Volcanic rocks*.

INTRUSION

Intrusion is the term applied to a body of rock that has invaded older rocks. It was originally restricted to, and is still most commonly used for, intrusive bodies of igneous origin. It is also used for masses of fragmentary rocks (pyroclastic rocks and breccias) infilling volcanic necks and fissures, and for sediments

that have been moved forcibly subsequent to their deposition.

Igneous intrusions are those bodies of igneous rock that have consolidated from magmas trapped under the surface of the earth. Depending on their size, they may be divided into *major* (e.g., batholith) and *minor* (e.g., dyke) intrusions, and on their mode of emplacement into *forceful* (forcible) and *passive* (permissive). Forceful intrusions divide into *concordant* and *discordant* where they either follow or transgress structures in the invaded rocks (e.g., dyke and sill, respectively).

An igneous intrusion is *simple* if it consists of the products of a single injection, *multiple* if it contains more than one injection of the same magma type, and *composite* if it is made up of a number of injections of different magma types. Simple intrusions that have undergone differentiation may show layering.

Salt domes are so named from the doming of strata over salt intrusions that have forms resembling igneous intrusions. On a small scale, mudstone may be intruded like salt, and it is not uncommon to find intrusion of sand into cracks and fissures owing to elastic expansion of pore fluids.

The following are related terms additional to those listed in the articles to which cross reference is given below.

Abyssal—related to an igneous intrusion, to its emplacement (abyssal injection), or to the rock of an intrusion that occurs at considerable depths as distinct from hypabyssal; abyssal rocks has been used as a term for rocks of major intrusions.

Apophysis—an offshoot of a large intrusion such as a tongue, vein, or dyke.

Basic border—the marginal part of an igneous intrusion that has a more basic composition than the interior part of the mass.

Bell-jar intrusion—an igneous intrusion like a bysmalith but with the strata of the sunken block domed and severely fractured.

Bysmalith—an intrusion with a roughly circular or plug-like form and bounded by faults.

Chonolith—an igneous intrusion whose form is so irregular and relations to intruded formations so complex that terms like sill, laccolith, or dyke are not applicable.

Country rock—the host rock of an igneous intrusion or mineral vein.

Ductolith—a horizontally disposed igneous intrusion that resembles a tear drop in cross-section.

Epiphysis—an apophysis detached from its source.

Feeder—a conduit through which magma passes during ascent, e.g., dyke, vent.

Floor—the country rock bordering the lower surface of an igneous intrusion.

Globulith—an intrusive mass, or group of associated masses, having almost concordant contacts and a globular or botryoidal shape.

Intrusion—the process of magmatic emplacement into preexisting rock as well as the body of rock resulting from such emplacement.

Intrusive (adjective)—relates to the processes of intrusion and the rock so formed (*note*: “intrusives” has been used, incorrectly, for intrusions or intrusive rocks).

Minor intrusion—an intrusion such as a dyke or sill rather than a major intrusion such as a batholith or lopolith.

Pluton—a deep-seated (plutonic) intrusion.

Rheomorphic intrusion—the injection of mobilized country rock into the igneous intrusion that caused the rheomorphism.

Ribbon injection—a tongue-like intrusion along the cleavage planes of a foliated rock.

Roof—the country rock bordering the upper surface of an intrusion.

Roof foundering (collapse)—the collapse of overlying rocks into underlying magma.

Roof pendant—a downward projection into an igneous intrusion of the country rock of its roof.

Schlieren arch—the result of flow layers forming the borders of an igneous intrusion, but absent in its interior.

Schlieren dome—the result of flow layers forming throughout an igneous intrusion and culminating in one central area.

Sheet—a tabular-shaped igneous intrusion; the term is commonly used for generally parallel-sided masses that are not clearly subhorizontal sills or near-vertical dykes.

Sole injection—an igneous intrusion emplaced along a thrust plane.

Sphenolith—a wedge-shaped intrusion that is partly concordant and partly discordant.

Stromatolith—a complex sill-like igneous intrusion that interfingers with sedimentary strata (cf., stromatite in **Migmatite**).

Stromoconolith—a conical or funnel-shaped layered igneous intrusion.

Subjacent—refers to an igneous intrusion (generally discordant) with no known floor and that presumably enlarges downward.

Subvolcano—a small, downward tapering intrusion occurring not far from the surface.

lith; *Migmatite*; *Phacolith*; *Plutonic rocks*; *Rheomorphism*; *Ring dyke*; *Sill*; *Volcanic neck*.

INTRUSION—LAYERED

Layered, or *stratiform intrusions* (banded differentiates) are bodies of igneous rocks that display a subdivision resembling stratification in sediments. This may take the form of mineral layering, or a regular compositional subdivision picked out by variations in modal mineralogy, or variations in whole-rock major- or trace-element chemistry. The layering reflects the behavior of growing crystals in a cooling magma chamber, and is influenced by such factors as magma viscosity, the relative densities of crystals and their parental liquid, convective movement in the magma itself, and the tendency of the liquid in a magma chamber to undergo a compositional, thermal, and density stratification prior to solidification.

Irvine (1982) considers a number of mechanisms by which crystals grow and accumulate within a cooling magma chamber.

1. *Gravitative crystal settling*—solids denser than the parent liquid settle out on the floor of the magma chamber in a manner analogous to clasts in a sediment.
2. *Flotation*—crystals that are less dense than the parental liquid accumulate against the roof of a cooling magma chamber; this is also a gravitative process.
3. *Nucleation*—crystals form against the walls, roof or floor of the chamber, initially by chilling of the magma against the country rock, but eventually by growth into a parent liquid replenished by convection currents.
4. *Accumulation*—aggregation of crystals held suspended in a parent liquid of high viscosity which precludes sinking or floating no matter what the density contrast.

Gravitative settling (effectively sedimentation in a magma) can operate in both intrusive and extrusive igneous rocks, and can affect both crystals separating from a parent melt, and segregating droplets of an immiscible liquid when the melt viscosity is low, density contrasts are high, and convective movements are not vigorous enough to maintain a thorough mixing. Such processes have been documented in thick, layered sills (e.g., Palisades sill, New Jersey, U.S.A.) and are invoked to explain sulfide accumulations in ultramafic lava flows (Usselman et al., 1979).

Double-diffusive convection permits the formation of a semistable stratification within a liquid magma chamber. The stability of such

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Cross-references: *Batholith*; *Breccia pipe*; *Cone sheets*; *Diapirism*; *Diatreme*; *Dyke (dike)*; *Emplacement—modes*; *Intrusion—layered*; *Laccolith*; *Lopo-*

layering depends on whether a magma chamber is filled by a single pulse of liquid, whether it is periodically replenished, or whether material is periodically vented. By using computer modeling techniques capable of handling large numbers of variables in concert, and considering both whole-rock and mineral major- and trace-element chemical data, these concepts have been applied to the petrogenesis of a number of well-known layered intrusions (Bushveld, Great Dyke of Zimbabwe, Stillwater, Skaergaard, Muskox—see Irvine, 1982 for references; Irvine, et al., 1983).

Layering in an intrusion may thus develop parallel to the floor, and horizontal with respect to the original attitude of the intrusion, or parallel to the roof and walls. Very large layered intrusions may contain all three, with the first type predominating, while smaller intrusions (dykes and plugs) tend to show the latter.

Types of layering

Rhythmic layering (small-scale layering) is present in the larger stratiform intrusions (Wager and Brown, 1968). There is a segregation of mineral phases into monomineralic or discrete assemblages (e.g., anorthosite, chromitite, troctolite) on a scale varying from a few centimeters to several meters, forming a stratification whose pattern is repeated through the thickness of an intrusion (Fig. 1). The recognition of rhythmic layering remains one of the principal criteria for identifying layered intrusions with the designation of different types of rhythmic layering based on mineralogical and textural contrasts between layers (Jackson, 1970; Irving and Dungan, 1980).

Rhythmic layering may be quite regular over great distances within an intrusion (Fig. 2), or there may be irregularities of composition, thickness, or mineralogy within one layer as traced through an intrusion. The similarities between these layers and sedimentary layering led to a terminology that was genetic (Wager et al., 1960; see **Cumulate textures**) and, although igneous sedimentation undoubtedly produced some cases of layering, this phenomenon can be interpreted in several ways (Maaløe, 1978; Irvine and Dungan, 1980; Irvine, 1982).

Cryptic layering is more subtle, and consists of a continuous or discontinuous change in mineral or whole-rock chemistry only expressed petrographically by changes in modal distribution of members of mineral solid-solution series, e.g., olivines, spinels, pyroxenes, and plagioclase feldspars. Some layered intrusions display a cryptic variation within a rhythmically layered sequence that includes virtually a complete

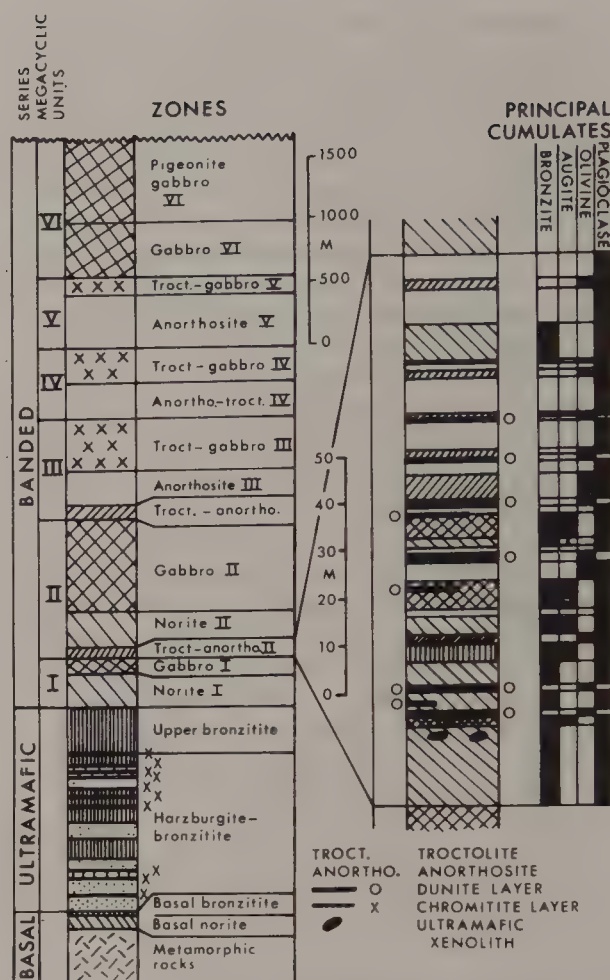


FIGURE 1. Detailed stratigraphy of the lower part of the Stillwater intrusion, Montana, U.S.A. (After Irvine et al., 1983.)

variation through the range of mineral solid solutions, e.g., Skaergaard intrusion. Such a layering is often regular enough to serve as a stratigraphy. Chemical characteristics such as these can be interpreted either as gravity settling from an evolving magma, or by crystal accumulation in a density stratified magma chamber (Irvine et al., 1983).

Examples of layered intrusions

Layered intrusions are known from many geological environments and of widely varying ages from Archaean to Tertiary. The *Skaergaard intrusion* (Tertiary) of E Greenland is the classic example and has been the basis of much modern structural and textural investigations of layered rocks. It is a small, well-exposed and intensively studied funnel-shaped body, displaying the products of marked fractional crystallization from gabbros to Fe-rich diorites in its main layered sequence, together with some residual acidic material. Rhythmic layering is particularly well developed, and



FIGURE 2. Bushveld Complex showing layering of pyroxenites of the Basal Zone (left) and norites and pyroxenites of the Critical Zone (center); Main Chromite Seam at base of hill to right of road; Olifants River in foreground.

certain distinctive features seem to indicate that strong magmatic convection currents operated spasmodically during deposition. Another important aspect of this intrusion is the occurrence of marginal facies (both on the walls and beneath the roof), which provide evidence of the initial magma composition and of other aspects of the early crystallization history not recorded in the layered rocks.

The *Bushveld Complex* of South Africa (ca. 2,000 Ma old) is best known for its vast size, apparently lopolithic form, and its great economic potential. Compositionally, it ranges from peridotites and pyroxenites in the lower part, through gabbros and anorthosites, to ferrrodiorites (and possibly residual acidic rocks) at the top. Rhythmic layering is locally well developed, especially in the vicinity of the chromite and magnetite horizons in the lower and upper parts of the intrusion, respectively (Fig. 2).

Other examples of ultrabasic–basic layered intrusions are well known for specific aspects of their structure and petrology. The *Stillwater intrusion* in Montana, U.S.A., has been steeply tilted since its emplacement, and the upper part appears to have been removed by erosion. Despite this, there is a thick (5,000 m) succession of peridotites, pyroxenites, and gabbros and anorthosites, which display prominent rhythmic layering locally, as well as distinctive textural features. The much smaller

Rhum intrusion (NW Scotland) crystallized beneath a Tertiary volcano, and shows cyclic repetition of major olivine-rich and feldspathic layers. Each unit (peridotite overlain by troctolite) probably represents the early crystal fraction extracted from a fresh batch of magma on its way to the surface. Spectacular rhythmic layering is also developed, including prominent slump structures, and a distinctive rock type (harrisitic peridotite) dominated by large branching olivines, characteristically oriented perpendicular to the layers. These olivines are interpreted as representing upward growth from the contemporary floor of the magma chamber during periods of nondeposition.

The *Great Dyke* of Zimbabwe is also a layered intrusion, despite its dyke-like shape (persistent over its length of 500 km). The rhythmic layering dips gently inwards from the steep walls. The bulk of the stratigraphic succession comprises a cyclic repetition of peridotites and pyroxenites, with important chromite layers occurring in most units. Gabbros are developed at the top of the succession at four centers along the length of the dyke, but nowhere are they very thick. The *Muskox intrusion* of NW territories, Canada, is also elongate in outcrop, with a dyke-like feeder beneath more normal layered peridotites and gabbros (funnel-shaped cross-section).

These examples of layered intrusions in stable cratonic areas are generally considered to have

crystallized from basic (basaltic) magma, even where they are predominantly ultrabasic or ultramafic in composition. Layered intrusions also occur in orogenic settings and at least three types are recognized, viz., layered gabbros in alpine ultramafic complexes or ophiolites, concentrically zoned ultramafic complexes and various hornblende-bearing layered gabbro-anorthosite complexes now found in deformed high-grade terrane or associated with major granitic batholiths. In addition, tectonic fragments of layered ultrabasic–basic masses occur in high-grade terrane. Metaperidotites show features similar to those of unmetamorphosed and undeformed masses, while the gabbro-norite-troctolite assemblage is represented by garnet–pyroxene rocks or retrogressed equivalents (Bowes et al., 1964).

Layered gabbroic bodies in ophiolites are considered to represent the products of the crystallization processes operative in the magma chambers beneath the spreading axes of mid-oceanic ridges or their equivalents in back-arc basins. Obduction of an ophiolite, and subsequent orogenesis both lead to the deformation and fragmentation of these oceanic layered intrusions. Nevertheless, in parts of the *Bay of Islands Complex*, Newfoundland, rhythmic and cryptic layering were developed and are preserved on an extensive scale. Layered ultramafic cumulates, layered gabbro, and massive gabbros, formed by the fractional crystallization of tholeiitic basaltic magma, are all present.

Concentrically zoned ultramafic complexes, many showing prominent layering, occur in certain orogenic belts, namely the Cordillera of SE Alaska, U.S.A., and the Ural Mountains, U.S.S.R. Feldspathic rocks are rare, and many of the ultramafic rocks are hornblende-bearing. Peridotite and pyroxenite dominate the assemblage. Their origin is rather obscure, but they may represent the deep levels of andesitic stratovolcanoes (Irvine, 1974).

Massive layered gabbro-anorthosites, typified by primary hornblende and highly calcic plagioclase, and occasional accumulations of Cr-spinel, occur primarily in high-grade Archaean gneiss terranes. The best documented example is the *Fiskanaesset Complex* in W Greenland (see **Anorthosite**). They are considered to have crystallized from tholeiitic magmas at high P_{H_2O} in the deep levels of orogenic belts. Other examples occur in the Archaean cratons of southern Africa and India, and similar intrusions are known from the Cretaceous of the Himalayas (Chilas Complex, Kohistan; see Coward et al., 1982) and related rocks may occur as mega-xenoliths in granitic batholiths in other Phanerozoic orogenic belts,

e.g., the Sierra Nevada batholith of California, U.S.A. Whether these examples constitute a distinct group is not clear.

Layering is not a phenomenon restricted to basic rocks; examples have been recorded in granites and syenites. In the Gardar province of SW Greenland layering is seen in both alkali granites and nepheline syenites; in the latter, the extreme differentiation combined with layering has produced spectacular effects (Upton, 1974; Parsons, 1979).

ADRIAN F. PARK

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Cross-references: *Anorthosite*; *Banding in igneous rocks*; *Bushveld Complex*; *Cumulate textures*; *Double-diffusive convection*; *Magma*; *Magmatic differentiation*; *Ophiolite*; *Skaergaard intrusion*.

INTRUSIVE IGNEOUS STRUCTURES—See **BATHOLITH**; **CONE SHEETS**; **DYKE (DIKE)**; **EMPLACEMENT—MODES**; **IGNEOUS INTRUSIONS—STRUCTURAL BEHAVIOR**; **INTRUSION**; **LACCOLITH**; **LOPOLITH**; **PHACOLITH**; **RING DYKE**; **SILL**; **VOLCANIC NECK**.

ISLAND ARCS—PETROLOGY

The volume of magmatic rocks being produced at the present day near convergent plate boundaries on island arcs and seismically active continental margins is huge, being second only in volume to the magmatic rocks produced at mid-ocean ridges (see **Mid-ocean ridge and ocean-floor petrology**). The diversity of igneous rock types developed in this setting is also very large, with the boninite, island-arc tholeiite, calc-alkali, high-K, and shoshonite associations (series) represented. All are in some way related to active plate convergence and subduction with the relationship between volcanism and seismic activity clearly demonstrated (Blot, 1981). Both the range of rock types represented and the problems involved in understanding their genesis are typified in the various island-arc structures around the Pacific area with the products of island-arc volcanism seen, for example, in the New Hebrides, Tonga, and Marianas arcs (Fig. 1). They are also seen in the Lesser Antilles in the Atlantic area, and the South Sandwich Islands in the Antarctic Ocean. Commonly there are marked changes in the igneous products during the evolution of an island arc, with extrusive members of the boninite and/or island-arc tholeiite series particularly important during the early stages (Gill, 1987) while later in the evolution of active continental margins, andesites and other volcanic and plutonic members of the calc-alkali series are prominent. There are also marked spatial variations in expression related to tectonic environment. These are illustrated by the dominance of basalts in the inter-oceanic island arcs (e.g., Marianas and Tonga arcs)

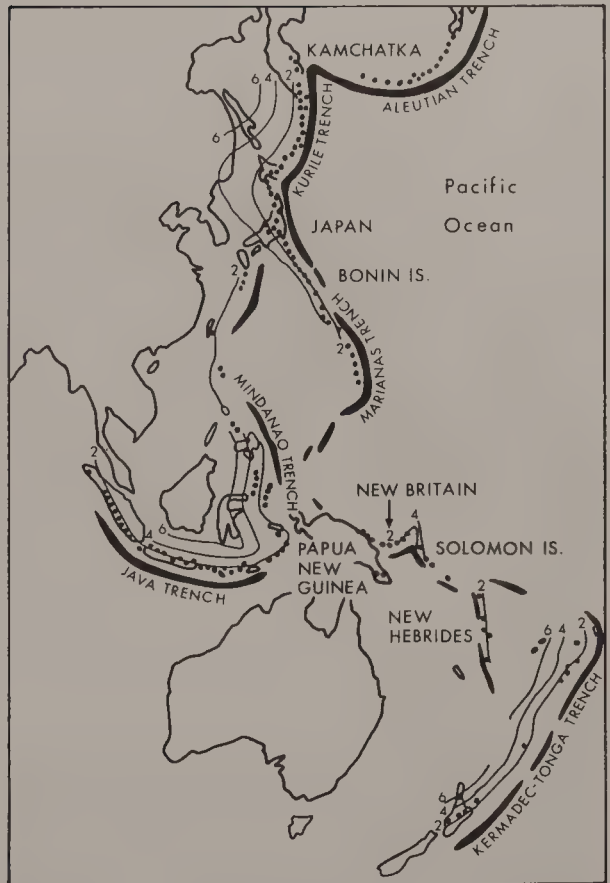


FIGURE 1. Distribution of trenches and volcanoes in relation to Benioff zones marked by 200-, 400-, and 600-km depth contours. (After Hall, 1987.)

evolved on oceanic crust with steep Benioff seismic zones compared to the dominance of deep-level plutonic activity and the silicic nature of the magma that reaches the surface in continental volcanic arcs (e.g., the Peruvian Andes) where the continental crust is thick and the Benioff zone has a low angle of inclination. Subduction in areas intermediate between these two extremes (e.g., in central America) is associated with the development of rocks of intermediate composition (usually andesites). Additional factors that contribute to the very complex pattern of igneous activity at convergent plate margins are variations in (1) the rates of plate convergence, (2) the composition of the material being subducted, (3) the thickness and composition of the buoyant plate margin, and (4) the physical properties and composition of the wedge of upper mantle material that occurs above the Benioff zone.

Morphology and rock associations

Typical island arc–trench systems have a characteristic morphology (Fig. 2) and characteristic rock associations. *Trench* areas have assemblages of deep ocean-floor type with deep-sea sediments and igneous and meta-

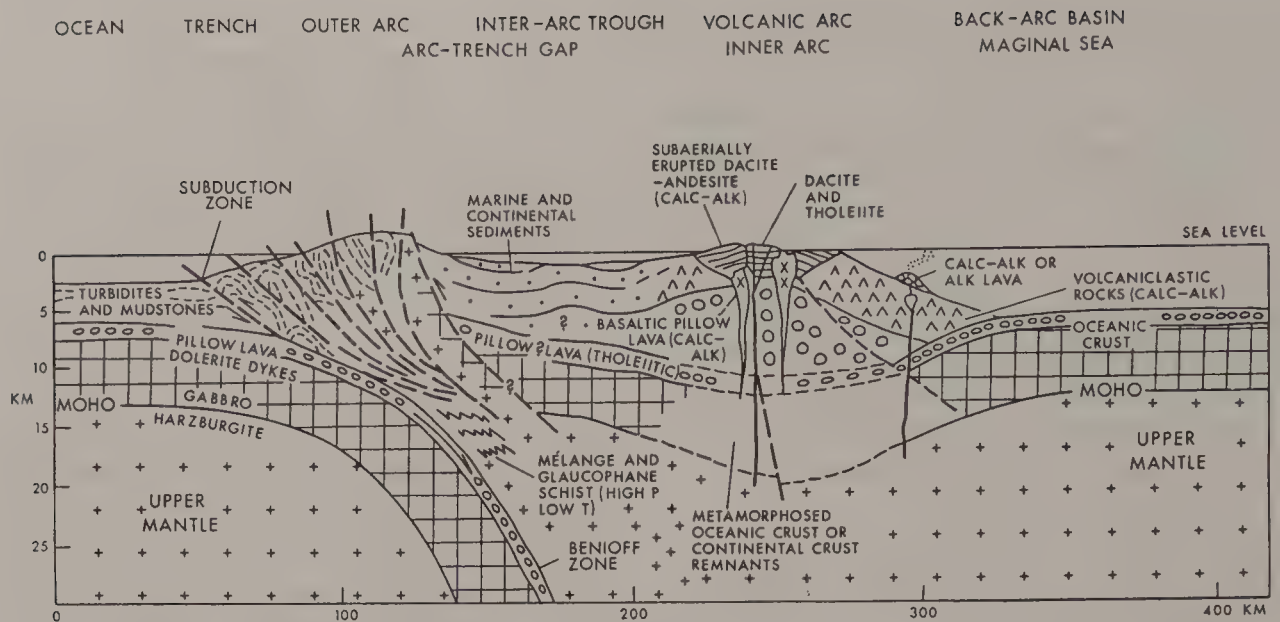


FIGURE 2. Diagrammatic cross-section of an arc-trench system. (After Garson and Mitchell, 1977.)

morphic rocks. Owing to the low heat flow, metamorphism that may occur in the deep part of the sedimentary pile is of high-pressure-low-temperature type. There is also the chaotic association of sediments with blocks and masses of basaltic and ultramafic rocks constituting *mélange*. The existence of negative gravity anomalies and the absence of large piles of sedimentary rocks in trenches suggest that large volumes of sediments and magmatic rocks were subducted down along the Benioff zone deep into the mantle. Trench rock associations can also be tectonically accreted to the existing island arcs as the result of obduction. There is no record of magmatic activity in trench areas with all volcanic rocks found there having come from mid-oceanic or island-arc centers or from seamount volcanic edifices. The *arc-trench gap* (also referred to as the *fore-arc* or *outer arc*, together with the *inter-arc trough*) is located between the oceanic trench and the volcanic front (the trenchward limit of the stratovolcanoes of the volcanic (inner) arc). Much of it is of nonvolcanic origin, metamorphic rocks are of the high-*P*-low-*T* type and there are both deep- and shallow-water sediments. The presence of ophiolitic rocks results from the tectonic accretion of oceanic crust. The depth of the inclined earthquake foci plane beneath the arc-trench gap is relatively shallow and usually is not > 100 km, but earthquake foci extend to at least 600 km. Positive gravity anomalies provide evidence of uplift, and heat flow is higher than over the trench. In some places there is submarine volcanic activity on submerged ridges adjacent to the trench (e.g., Kamchatka). The *volcanic arc* (*inner arc*) is

manifested by the appearance of volcanic activity above sea level. It is characterized by positive gravity anomalies, the presence of voluminous volcanic rocks, shallow-water sediments and low-*P* metamorphic rocks. In Japan, the Kuroko Zn-Cu-Pb mineral deposits lie within the arc tholeiite zone at a depth of ca. 150 km above the present-day depth of the Benioff zone (Fig. 3). The *back-arc region*

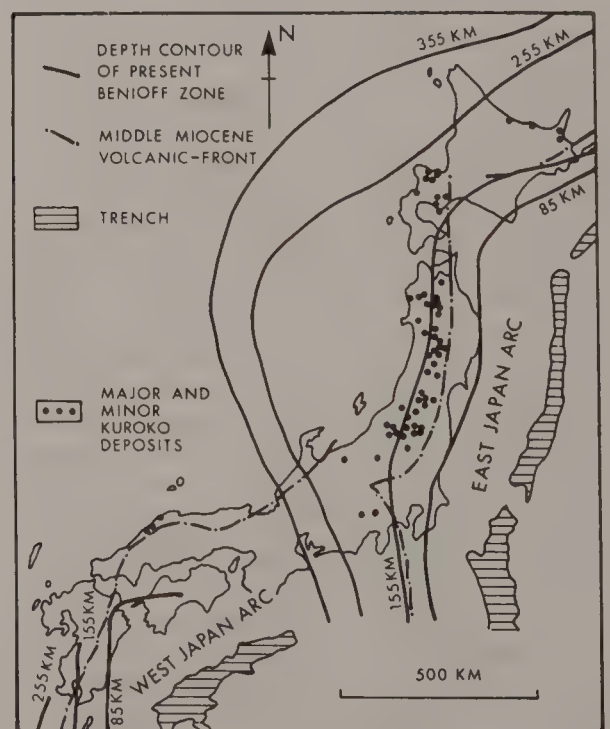


FIGURE 3. Distribution of Kuroko ore deposits in relation to depth of Benioff zone and position of the volcanic front. (After Sato, 1977.)

(*marginal sea*) that lies behind the volcanic arc may also contain the products of volcanic activity, although these normally differ in trace-element and isotopic composition from the magmatic products of the volcanic arcs (Gill, 1981). In Vanuatu, New Hebrides, basaltic rocks dominate both back-arc and volcanic arc rock assemblages, while in the Tonga region the extrusions in an actively spreading back-arc region are submarine basalts, while andesitic stratovolcanoes dominate the volcanic arc.

Major- and trace-element variations

One unifying feature that links the great variety of magmatic products in island arcs is the commonly recorded systematic relationship in abundances of incompatible elements of rocks with similar SiO_2 content to depth to the Benioff zone. This is particularly well shown by potassium (Fig. 4) and reflected by the K_2O vs. SiO_2 plots for the various volcanic series that represent the corresponding overall lateral variation away from the volcanic front (Fig. 5). There is corresponding increase in the $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratio and decrease in FeO/MgO (Jakeš and White, 1972). These features and the spatial distribution of the various types of volcanic rocks (e.g., the basaltic rocks of Japan—Aramaki and Ui, 1982) are consistent with the magmatic activity being triggered by processes within, or associated with, subducted slabs of oceanic lithosphere: the release of water associated with dehydration of hydrous phases in the descending old oceanic crust (with ca. 5% H_2O) is considered to play a key role. However, radiogenic isotope and trace-element criteria that have been developed for the recognition of contributions from the subduction zone tend not to correlate with suites of

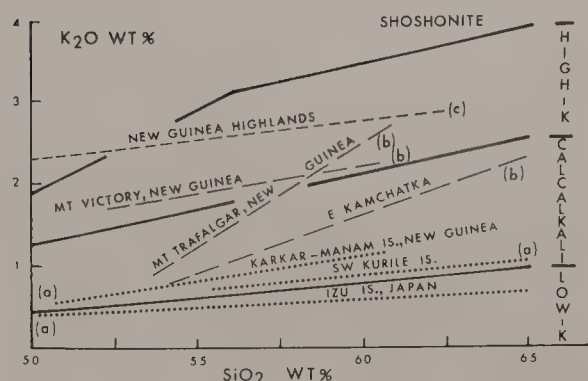


FIGURE 5. K_2O vs. SiO_2 in volcanic rocks of island arcs showing subdivision into low-K (boninite and arc-tholeiite), calc-alkali, high-K, and shoshonite series and examples of (a) tholeiite, (b) calc-alkali, and (c) shoshonite series. (After Jakeš and White, 1972, and Middlemost, 1985.)

subduction zone rocks (cf., Ellam and Hawkesworth, 1988).

The volcanic rocks of island arcs are typically high in Al_2O_3 , irrespective of SiO_2 content. In typical island-arc tholeiites (including tholeiitic andesites and dacites) the range is 14–17.5%; in the calc-alkaline rocks it is 16–17.5% over most of the SiO_2 range, while in the high-K and shoshonitic associations it is 14–19%. $\text{Fe}_2\text{O}_3/\text{FeO}$ increases away from the trench. Of the trace elements, large cations (Rb, Ba, Sr, Pb) show the largest variations, e.g., K/Rb decreases from 1,000 in the island-arc tholeiite series to 250 in the shoshonite series, while Ba in basaltic members increases from 75 p.p.m. in tholeiites to 1,000 p.p.m. in shoshonites. The bulk content of rare-earth elements is usually low (<100 p.p.m.), with island-arc tholeiites (and tholeiites of back-arc basins) showing flat chondritic-type patterns, while the calc-alkali, high-K, and shoshonite series show strongly fractionated patterns (Fig. 6). At the same SiO_2 level there is an increase of the large, high-valency cations, such as Th and U, away

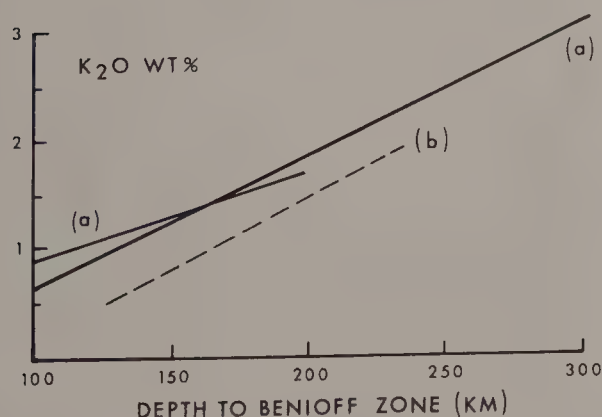


FIGURE 4. K_2O content of andesitic lavas at constant SiO_2 content of 55% vs. depth to Benioff zone beneath island arcs. (a) Averaged results from a number of island arcs; (b) Sunda arc, Java, and Bali, Indonesia. (After Middlemost, 1985.)

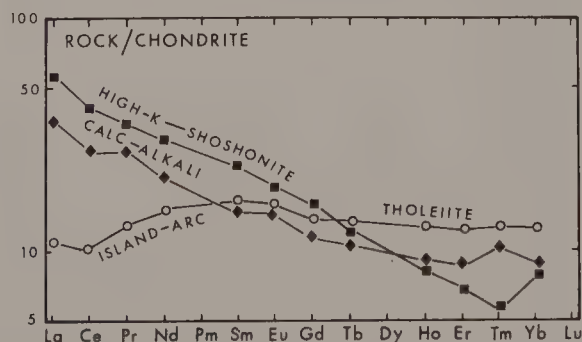


FIGURE 6. Distribution of rare-earth elements to chondritic abundances in rocks of island-arc volcanic associations.

from the arc-trench gap (Jakeš and White, 1972).

Volcanic series

Boninite series. In addition to the more voluminous rocks of the island-arc tholeiite series, the deepest parts of volcanic sequences of several island arcs of the western Pacific region contain high-Mg and low-Ti basalts, basaltic andesites, and dacites of the boninite series, the Marianas arc-trench slope (Papua-New Guinea) and the Bonin island arc (ca. 1,000 km S of Tokyo, Japan) being examples (Middlemost, 1985).

Island-arc tholeiite series. Rocks of this series are relatively common during the early stages of island arc development. Together with the boninite series they are often referred to as the low-K series, as they are characterized by low K_2O values. They plot as being tholeiitic on a FeO^*/MgO vs. SiO_2 diagram, display rapid Fe-enrichment, and are usually restricted to the volcanic front of plate boundaries characterized by convergence rates of >7 cm/year (Gill, 1981). They are most voluminous in the northwestern Pacific region but make up only a small proportion in the northeastern Pacific region and a comparable proportion in the Aeolian and Aegean arcs of the Mediterranean region (Table 1).

Calc-alkali series. While there is no generally accepted clear definition of this series, orogenic andesite is a characteristic member. The series consists of subalkalic, silica-oversaturated rocks that tend to contain more Al_2O_3 than normal rocks of the tholeiitic series, and their intermediate members do not normally show the effects of significant Fe-enrichment. They predominate in the northeastern Pacific region and are abundant in both the Andean region of S America and the Aeolian and Aegean arcs of the Mediterranean region. In the volcanic assemblages of New Hebrides and New Britain, there are rocks showing gradations between the tholeiite and calc-alkali series (cf., Coulon and Thorpe, 1981).

High-K and shoshonite series. Intermediate volcanic rocks of these series are uncommon or absent in the northeastern Pacific region. In the southwestern Pacific region and the Andean region of S America they are much more common, while in the Aeolian and Aegean arcs of the Mediterranean region they constitute $>40\%$ of the assemblage. They characteristically occur in areas where the Benioff zone is at its deepest. Associated rhyolites are thought to represent (or largely so) the products of crustal anatexis rather than magma derived from much greater depths.

TABLE 1. Abundance of low-K, calc-alkali, high-K, and shoshonite series in intermediate volcanic rocks in present day orogenic areas.

	A ^a	B ^b
Southwestern Pacific region		
Shoshonite series	6.9	2.0
High-K series	21.8	25.6
Calc-alkali series	43.7	57.8
Low-K series	27.5	14.6
Northwestern Pacific region		
Shoshonite series	1.2	1.6
High-K series	11.5	12.6
Calc-alkali series	57.1	62.8
Low-K series	30.1	23.0
Northeastern Pacific region		
Shoshonite series	0.0	0.0
High-K series	1.2	8.8
Calc-alkali series	94.7	87.7
Low-K series	4.1	3.5
Andean region, S America		
Shoshonite series	17.6	4.9
High-K series	21.6	61.8
Calc-alkali series	50.0	32.6
Low-K series	10.8	0.7
Aeolian and Aegean arcs, Mediterranean region		
Shoshonite series	26.0	21.4
High-K series	16.8	37.8
Calc-alkali series	49.6	38.7
Low-K series	7.6	2.1

Source: after Ewart (1982) and Middlemost (1985).

^a A Relative abundance (%) of volcanic rocks containing 52–60% SiO_2 .

^b B Relative abundance (%) of volcanic rocks containing 56–63% SiO_2 .

Petrogenesis

Many authors have favored the development of basaltic magma in subduction zones as the result of the melting of subducted lithosphere (oceanic crust), with Marsh (1982) matching the commonest type of lava in the Aleutian Is. with the product of 60% melting of eclogite. On the other hand, many authors have favored derivation from peridotite of the adjacent mantle, which is the presumed source of basaltic magma in all other tectonic environments (cf., Ringwood, 1974). There has also been debate concerning the presence of sedimentary material in source regions, but Pb and Be isotopic compositions demonstrate a definite, though small, contribution to arc magmas (e.g., Barreiro, 1983; Tera et al., 1986; White and Dupré 1986). Multiple sources, including oceanic crust, mantle, and continental crust have also been proposed (e.g., Hickey et al., 1986).

Experimental studies have concentrated on

(1) the partial melting of mantle peridotite in the presence of water, (2) the partial melting of eclogite facies basaltic material, (3) the partial melting of amphibolite facies basaltic material at shallow depths, and (4) amphibole-controlled fractionation of basaltic material formed in the subducted crust or the overlying mantle. Concepts related specifically to the genesis of andesites include (a) melting of basic rocks within the continental crust, (b) differentiation of basaltic magma associated with either high-pressure fractionation ("eclogite fractionation") or low-pressure fractionation, (c) contamination, and (d) magma mixing (Hall, 1987).

Resolution of these problems has come from integrated studies of Nd and Sr isotopes and of major and trace elements (Perfit et al., 1980), with Rogers and Hawkesworth (1989) showing the mantle wedge to be the principal site of magma development beneath the North Chilean Andes, with variations in Nd and Sr signatures explained by variations in the nature of the wedge as it migrates. Triggering of the melting in the mantle wedge is caused by the influx of fluids associated with the dewatering of the downgoing slab as the result of amphibole breakdown (Plank and Langmuir, 1988), something that is related to the depth of the subducted lithosphere. This evidence for arc-magmatism being related to hydrous melting of a peridotite mantle beneath the volcanic fronts is consistent with the evidence from experimental petrology presented by Wyllie (1984), as is the similarity in major-element composition of mid-ocean ridge and arc basalts. The extent of melting in the mantle is related to the thickness of the mantle wedge beneath the crust of the volcanic arc, which in turn depends on the thickness of the overlying crust. The apparent lack of correspondence in some geochemical signatures with certain petrologic types, and the presence of some apparently contradictory geochemical trends, can be explained as the result of the operation of migrating magmatism that sampled a mantle wedge of varying composition, together with the preferential scavenging of certain elements by the hydrous fluids liberated into the wedge. The fractionation of the resultant magmas in the complex and varied geotectonic environment leads to the great variety and complexity of the island-arc rock assemblage.

D. R. BOWES

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Cross-references: *Andesite and dacite; Basalt; Basic igneous rocks; Calc-alkaline rocks; Circum-Pacific plutonic and volcanic activity; Eclogite; Experimental petrology; Igneous rocks—tectonic setting; Intermediate igneous rocks; Magma; Magmatic differentiation; Metamorphic facies series; Mid-ocean ridge and ocean-floor petrology; Oceanic crust; Peridotite; Upper mantle—experimental petrology; Volcanoes and volcanism.*

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KHONDALITE

Sillimanite- and garnet-rich siliceous schists, gneisses, and granulites, together with minor proportions of garnetiferous quartzite and of a white rock resembling limestone, that make up a considerable part of the Eastern Ghats terrane of India, were referred to by Walker (1902) "for the sake of brevity of nomenclature and as a local convenience", as *khondalite*. The name was after the Khond tribe that inhabits the Baud-Khondmals (now Phulbani) region of Orissa. They are associated with 2-pyroxene granulite, charnockite, and anorthosite, and similar rocks occur in many other parts of Peninsular India and in Sri Lanka and central Burma (Krishnan, 1968).

The major phases—quartz, garnet, and sillimanite—vary considerably in modal proportions, with K-feldspar and plagioclase major constituents in some rocks. Grain size is generally 2–3 mm; commonly there is a foliation subparallel to lithological layering, but in some rocks the texture is granulose. Polyphase metamorphic mineral growth is shown with the peak of metamorphism being $>600^{\circ}\text{C}$. Some leucosome–melanosome banding may represent partial melt and restite.

An average chemical composition for the khondalites of Orissa shows that $\text{SiO}_2 + \text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3 + \text{FeO}$ constitute 95% of the rock with $\text{Fe}_2\text{O}_3 > \text{FeO}$. The grouping close to the A–F sideline on an ACF plot (Fig. 1, left) is even tighter than that for continental clays, while on a $\text{SiO}_2\text{--Al}_2\text{O}_3\text{--Fe}_{\text{TOT}}$ plot (Fig. 1, right) the overall field for the khondalites is similar to that for present-day deeply weathered soil profiles in those parts of Brazil where weathering has not been extreme. This correspondence also extends to enrichment, stability, or depletion, compared to crustal averages for Ce, Co, Cu, Ga, La, Ni, Nb, Th, U, Y, Zn, and Zr. These data are consistent with the quartz–sillimanite–garnet part of the khondalite assemblage in Orissa, and presumably in adjacent regions, being granulite facies–upper amphibolite facies metamorphic equivalents of a deeply weathered soil profile (or soil profiles). The associated siliceous and calcareous units (quartzites and calc-silicate granulites containing calcite, diopside, wollastonite, scapolite, and plagioclase) have geochemical characteristics consistent with being high-grade metamorphic silcretes and calcretes, respectively (Dash et al., 1987).

In China the term khondalite has been used for high-grade metamorphosed aluminous rocks, but elsewhere outside India descriptive terms such as sillimanite–garnet gneiss (or schist) have generally been preferred. It has been used for rocks occurring in the granulite belt of N Finland for which a sedimentary parentage of a graywacke–shale sequence is proposed (Barbey et al., 1982).

B. DASH

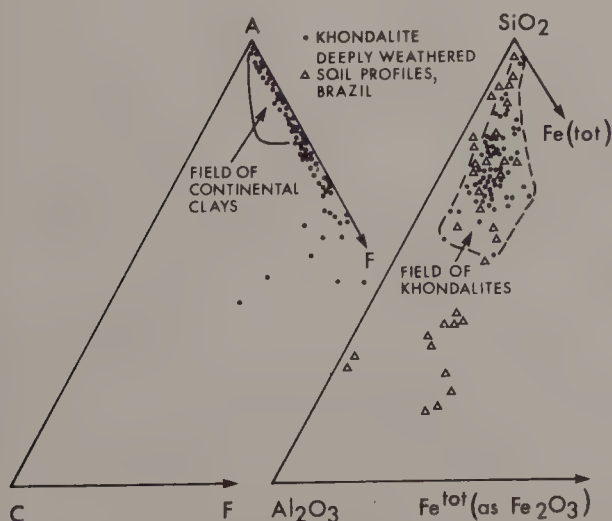


FIGURE 1. Plots and fields of composition of khondalites from Orissa, India, and comparisons with (left) continental clays and (right) present-day deeply weathered soil profiles, Brazil.

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Cross-references: *Gneiss*; *Granulite*; *Granulite facies*; *Schist*.

KIMBERLITE

Kimberlite, as originally described by Lewis (1887, see Dawson, 1980) was a serpentinized, ultrabasic, phlogopite-bearing, diamondiferous volcanic breccia found at the Kimberley diamond mine, S Africa. The term blue ground refers to unoxidized slate-blue or blue-green kimberlite that occurs below the superficial oxidized zone.

Since its original description, “kimberlite” has been extended to nonbrecciated rocks of similar mineralogy to that from the type locality but lacking diamond, and also to others that, although petrographically different from Kimberley Mine kimberlite, do contain diamonds. Indeed, Wagner (1914, see Dawson, 1980) described two distinct types of kimberlite—“basaltic” and lamprophyric varieties. Although it is now appreciated that textural differences may be due to different modes of emplacement (Dawson and Hawthorne, 1970, see Dawson, 1980), the presence or absence of distinctive phases, together with wide ranges in the proportions of other phases, suggest that the

term kimberlite is best applied to a *group* of closely related rocks, rather than to a single, petrographically specific rock-type (Dawson, 1987).

Although it is rare, kimberlite is important because (1) it is the main primary source of diamond; (2) it contains blocks of a wide variety of rocks derived from the upper mantle, and has carried out a more extensive sampling of the upper mantle than any other type of magmatic activity; and (3) it contains a much higher concentration of “incompatible” trace elements than other ultramafic rocks, which focuses attention upon a source for these elements within the upper mantle and the processes concentrating them in kimberlite.

Kimberlites are confined mainly to ancient shield areas, or areas underlain by shields (Fig. 1). They are anorogenic intrusions, being found on ancient, upwarped shield areas (e.g., S African plateau, Siberian platform) into which they were injected along broad elongate zones of weakness during epeirogenic movements, and in the tectonic zones on the margins of major horst-blocks. On a local scale, they occur on domes and on elongate swell structures, some of which were sufficiently dilated to become graben-faulted.

Kimberlite most commonly occupies small diatremes, sills, and dykes (Wagner, 1914; Williams, 1932, see Dawson, 1980). However, a kimberlite ring dyke has been reported from Sierra Leone and kimberlite lavas are known in Tanzania. In areas such as S Africa where the diatremes have been mined out sufficiently



FIGURE 1. World distribution of kimberlites and diamonds.

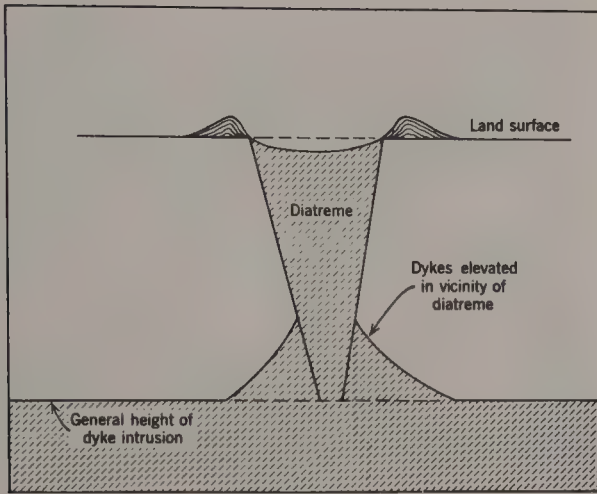


FIGURE 2. Diagrammatic section showing the depth relationships between a kimberlite diatreme and its feeder dyke.

deeply, or in deeply dissected areas, it can be shown that zones of diatremes are the surface expressions of kimberlite dykes (Fig. 2). These dykes infill deep, fundamental fractures that often cut across the deepest known basement structures. The kimberlite in these intrusions varies in appearance from a fragmental tuffaceous or agglomeratic rock, through fragmented material cemented by massive magmatic kimberlite (kimberlite-breccia), to a massive rock with magmatic textures; the fragmental varieties predominate in the diatremes, whilst the others are characteristic of sills, dykes, and the deeper parts of diatremes. These two contrasting types have been classified as *diatreme-facies kimberlite* and *hypabyssal facies kimberlite*, respectively. Flow textures and magmatic sedimentary textures are found in the massive variety (Dawson, 1980). This contrast in lithology can be directly attributed to their differing but intimately associated modes of emplacement. Whereas the hypabyssal facies rock results from consolidation of the magma at depths, the diatreme-facies rocks formed when kimberlites broke through to the surface, following which the explosion vent was enlarged and infilled by fluidized fragmental kimberlite. The surface expression was probably a crater with a tuff ring of fragmental kimberlite. If the crater contained a lake, tuff-ring material washed back into the crater would form kimberlitic shales and sandstones, such as those preserved in the noneroded diatremes of Nwadi, Tanzania, and Orapa, Botswana.

Kimberlite magmatism has occurred several times, one major episode of intrusion being in the Cretaceous. The accumulated evidence from both stratigraphic sources and radiometric dating is that certain geographical areas, usually coinciding with old cratons, have been subjected

to intermittent kimberlite intrusion throughout geologic time. The oldest kimberlites in Liberia may be 2.7 Ga old, whereas those of the Igwisi Hills, Tanzania are Recent.

A characteristic feature of kimberlite is its inequigranular texture—relatively large rock or mineral fragments are set in a fine-grained matrix. Four types of rock-fragment or xenoliths are found in kimberlite intrusions: (1) those derived from formations present in the area at the time of intrusion, but which have since been eroded from the area; (2) fragments of the immediate wall rocks; (3) xenoliths derived from underlying basement terranes or the deep crust (e.g., gneisses, banded ironstones, granulite); and (4) xenoliths of peridotitic and eclogitic rocks, the density and mineralogy of which are consistent with an origin in the upper mantle. The inclusions may be regarded as a composite section through the Earth from the point of kimberlite origin within the mantle to the Earth's surface.

Apart from rock fragments, the kimberlite is studded with macrocrysts that give the rock its “porphyritic” appearance (Fig. 3). The essential feature of kimberlite petrography and mineralogy is that kimberlite comprises a set of high-temperature macrocrysts (both xenocrysts

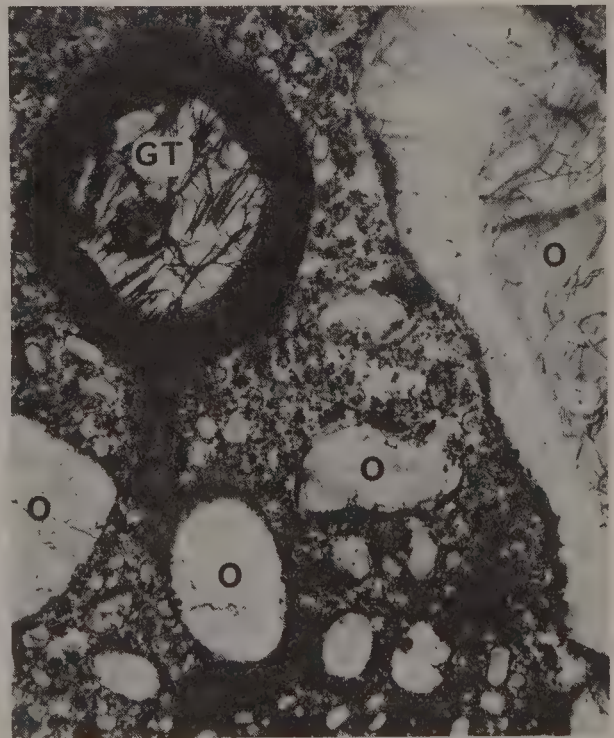


FIGURE 3. Photomicrograph of kimberlite showing megacrysts of garnet (GT) and partially serpentinized olivine (O) set in a matrix of smaller olivines, calcite, perovskite, and magnetite; the garnet is surrounded by a reaction rim of very fine-grained hydrous minerals; Koffiefontein Mine, S Africa; PPL, $\times 8$.

and phenocrysts) embedded in a finer-grained, low-temperature, volatile-rich matrix with which the macrocrysts have interacted. Macrocrysts of forsterite, enstatite, Cr-diopside, pyrope, and phlogopite were derived from fragmented lherzolite xenoliths, whereas more calcic garnet and omphacite macrocrysts (together with rare kyanite and corundum) came from fragmented eclogite xenoliths (Dawson and Stephens, 1975; Stephens and Dawson, 1977, both cited in Dawson, 1980). In addition, there are large megacrysts (up to 20 cm) of Ti-pyrope, Ti-phlogopite, olivine (Fo₈₂₋₉₀), bronzite, and ilmenite. They are too large to come from xenoliths, are chemically distinct from xenolith phases, and are interpreted as high-temperature phenocrysts (Gurney et al., 1979). Ilmenite may be intergrown with, or included in, the other megacryst phases. A feature of kimberlite is that the xenoliths, xenocrysts, and macrocrysts show some degree of interaction with the kimberlite matrix; the lherzolite and eclogite xenoliths have been affected by varying degrees of serpentinization, calcitization, phlogopitization, or silicification.

The matrix consists of varying proportions of serpentine, calcite and/or dolomite, spinel, perovskite, apatite, ilmenite, diopside, monticellite, phlogopite, and small euhedral crystals of forsterite (Dawson, 1980). The kimberlite is named according to the dominant groundmass phase, e.g., diopside kimberlite, calcite kimberlite. The groundmass mica is chemically distinct from xenocrysts or megacrysts. It contains inclusions of other matrix phases, such as spinel (Fig. 4), and often has anomite rims; it contrasts with the megacryst phlogopite, which is often corroded and distorted.

Zircon is a common accessory mineral; rarer recorded matrix minerals are titan-olivine, titan-clinohumite, danburite, datolite, pyrochlore, and periclase. There is also a wide variety of secondary minerals; some, such as serpentine or perovskite, result from macrocryst-matrix reactions, whereas others are primary low-temperature hydrothermal minerals. The latter group includes minerals of the zeolite, amphibole, and hydromica groups together with low-temperature sulfates and sulfides. Diamond, when present, is a rare accessory mineral. Overall, kimberlites tend to have two generations of olivine, diopside, phlogopite, and ilmenite (phenocrystal megacrysts and matrix). This feature, together with their texture and high volatile content, suggests kimberlites should be categorized as a variety of lamprophyre.

Kimberlite is an undersaturated ultrabasic rock, but within these terms it varies widely in composition, this being partly due to its hybrid



FIGURE 4. Photomicrograph of micaceous kimberlite showing macrocrysts of serpentinized olivine and a phenocryst of brown mica (M), with a dark oxidized rim, set in a fine-grained matrix containing randomly oriented phlogopite mica; Hololo dyke, Lesotho; PPL, $\times 38$.

make-up (Fig. 5). Dawson (1967) has shown that both the major and trace elements in kimberlites can be divided into two groups: (1) elements present in amounts similar to those in peridotite or lherzolite, and (2) those occurring in much higher concentrations than in ultrabasic rocks. It is variations in the proportions of these two groups of elements that give kimberlites their wide range in chemical composition. The second group of elements is concentrated within the matrix. Relative to other peridotitic rocks, kimberlites are enriched in the following major oxides: TiO₂, Al₂O₃, Fe₂O₃, Cr₂O₃, MnO, CaO, K₂O, P₂O₅, H₂O, and CO₂. They are, however, poorer in SiO₂ and MgO. In addition, they have K₂O > Na₂O and lower Mg/Fe ratios (Table 1). As well, kimberlite, and hence its matrix, is relatively enriched in the following minor elements: Li, B, F, Sc, V, Cu, Ga, Rb, Sr, Y, Zr, Nb, Cs, Ba, La, Ta, Pb, Th, and U, the relative enrichment being underlined by low Al/Ca, K/Rb, Nb/Ta and Ca/Sr ratios (Dawson, 1974, see Dawson, 1980). Some of these elements, notably Sr, Y, Zr, Nb, Ba, La, and Pb, are concentrated in amounts similar to those in carbonatites. The

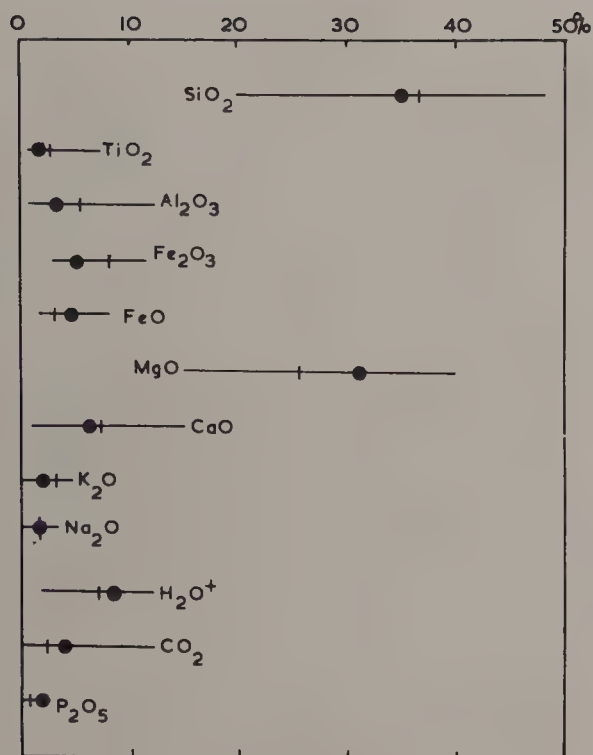


FIGURE 5. Variation in oxide wt.% in kimberlites; circles and bars on the variation lines represent values for the average kimberlite and micaceous kimberlite, respectively (see Table 1).

TABLE 1. Major element content of garnet peridotite and kimberlite

	1	2	3
SiO ₂	46.53	35.2	31.1
TiO ₂	0.10	2.32	2.03
Al ₂ O ₃	1.84	4.4	4.9
FeO (total)	6.10	9.8	10.5
MnO	0.11	0.11	0.10
MgO	41.98	27.9	23.9
CaO	1.47	7.6	10.6
Na ₂ O	0.16	0.32	9.31
K ₂ O	0.15	0.98	2.1
H ₂ O +	2.90	7.4	5.9
P ₂ O ₅	0.02	0.72	0.66
CO ₂	0.20	3.3	7.1
K/Na	1.0	3.4	7.5
Mg/Fe	4.8	2.2	1.8

1. Average of nine garnet peridotite xenoliths in S African kimberlites: the small proportions of K₂O, H₂O, P₂O₅, and CO₂ are probably derived from secondary minerals caused by percolation of solutions from the host kimberlite.

2. Average kimberlite (Dawson, 1967).

3. Average micaceous kimberlite (Dawson, 1967).

rare-earth element patterns have high light rare-earth and low heavy rare-earth levels. The ⁸⁷Sr/⁸⁶Sr values of 0.7038–0.7040 for kimberlite groundmass (calcite) are also within the range for igneous carbonates. The range in chemical composition of kimberlites reflects not only the amount of xenolithic material but also wide differences in the presence and proportions of different matrix minerals. Some features, such as high calcite and apatite, are more enhanced in hypabyssal kimberlites, probably reflecting quiet intrusion with volatile retention. Other features, such as the high phlogopite content of some matrices compared with virtual absence in others, are not yet understood.

The genesis of kimberlite is a matter of considerable debate (Dawson, 1980): three of the more modern hypotheses, all starting with various varieties of parental garnet peridotite, are discussed below. One of the most important features of kimberlite, its very limited volume, appears to preclude the first hypothesis, which proposes that kimberlite is the end product of a zone-refining process in the upper mantle (Harris and Middlemost, 1969, see Dawson, 1980), a process that might be expected to produce considerable quantities of magma. Indeed, the limited volume of kimberlite suggests that it is either the final product of a larger magmatic cycle or the product of incipient melting in the upper mantle. The former process, which has been invoked in the "residual" hypothesis, starts with a garnet lherzolite parent (pyropic garnet + Cr-diopside + enstatite + olivine); partial melting involving the garnet–diopside fraction at depths of 80–100 km is thought to produce a liquid with the composition of picrite basalt, the more refractory olivine–enstatite portion remaining as dunite or peridotite. Fractional crystallization at high pressures of the picrite basalt is considered to give rise to bimineraleclogite cumulates, and a series of silica-poor alkalic residual liquids that have the geochemical characteristics of kimberlites and some other rare ultrabasic potassic lavas.

The alternative process is termed the "incipient melting" hypothesis. In this case it would be necessary to have some low-melting phase containing the volatile and other "incompatible" elements. Dolomite or calcite are probably present (Wyllie, 1979, see Dawson, 1980) and phlogopite, amphibole, and titan-clinohumite are all hydrous phases that have been proposed as existing in the upper mantle. However, on the basis of the phlogopite xenocrysts present in kimberlite and current experimental results on high-pressure hydrous phases, phlogopite appears the most likely hydrous and potassic phase deep in the upper

mantle where kimberlite originates. In the incipient melting hypothesis it is envisaged that the carbonates and mica melt first, followed by a small amount of the Cr-diopside. The liquid formed by the incipient fusion would be rich in K_2O , Na_2O , TiO_2 , Al_2O_3 , CaO , FeO , Cr_2O_3 , and H_2O compared with the bulk parental peridotite, and would be a medium from which the olivines, pyroxenes, magnetites, perovskites, chromites, magnesian ilmenite, and garnets in kimberlites could crystallize. This melt, modified by crystallization of these phases, would approximate to the kimberlite matrix. This matrix, with its crystallized phases and varying proportions of blocks and fragments of lherzolite and eclogite would finally consolidate as kimberlite, after incorporation of some crustal xenoliths and interaction with high-level groundwater (Dawson, 1980). Whether the various varieties of kimberlite result from fractionation of a single parent, or whether they reflect real differences in magma generation in the upper mantle, is not yet clear. In addition, the relationship of the kimberlite group to similar rock-types, such as alnöite and leucite lamproite (Atkinson et al., 1986; Jaques et al., 1982) is yet to be clarified. However, recent isotope studies suggest that, unlike kimberlite, micaceous kimberlites and lamproites derive from old, metasomatized mantle (Dawson, 1987).

J. B. DAWSON

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- Cross-references: *Breccia pipe*; *Carbonatite*; *Diatreme*; *Eclogite*; *Igneous rocks—tectonic setting*; *Lamproite*; *Lamprophyre*; *Mafic and ultramafic inclusions in igneous rocks*; *Ultramafic lavas*; *Ultramafic rocks*; *Upper mantle—experimental petrology*; *Xenolith*.

* Contains extensive list of references.

L

LACCOLITH

A *laccolith* is a lensoid igneous intrusion that is concordant with the stratification or other type of banding in the host rock. It is usually circular in plan, <5 km across and rises from a flat base to form a low dome <100 m high, with a volume of 1–10 km³. Those of Utah, Wyoming, and Montana, U.S.A., belong to

volcanic environments and most can be related to a local subaerial volcanic cone or vent (Fig. 1; Eardley, 1962). Typically, several laccoliths surround the cone and they may represent more than one phase of magmatic intrusion.

Doming associated with laccolith formation is due to magmatic pressure unassisted by tectonic stress. The intrusion begins as a small sill with a stable roof but, as the intrusion extends, the roof becomes relatively weaker and yields by doming or by fracture on annular faults. Hence, laccoliths constitute a special case of sills favored by thinness, weakness of roof rocks, and viscosity of the intruding magma, which is commonly acidic or intermediate in character. In other cases it crystallizes shonkinite (Fig. 2) or teschnite or essexite. Some laccoliths are compositionally and texturally zoned horizontally or concentrically owing to crystal fractionation of the magma in situ, or to pulsative intrusion from the central vent, where differentiation may also be occurring, or to flow differentiation. As in the case of sills, repeated intrusion may result in simple, multiple, and composite laccoliths. Those having several parts that are separated from each other by thick layers of country rock, but were formed by a single intrusion, are referred to as *compound laccoliths*.

A special group of laccoliths arises from the submarine intrusion of spilitic lavas into soft

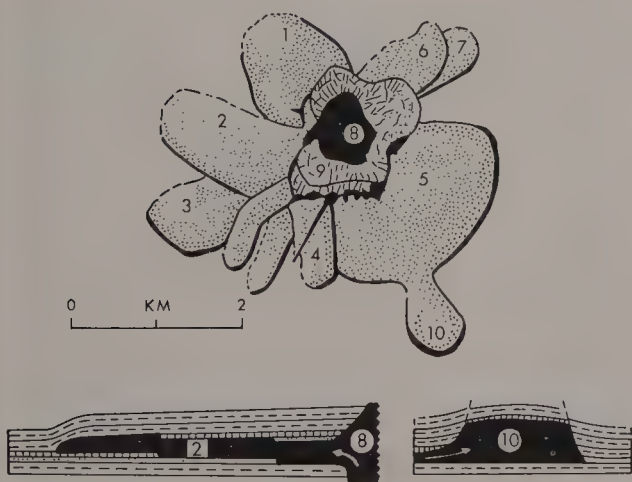


FIGURE 1. Model of igneous core of Mt. Ellen, Utah, U.S.A., stripped of cover to show intrusive stock (8), peripheral shatter zone (9) and satellite laccoliths 1–7 and 10. (After Eardley, 1962.)



FIGURE 2. Shonkin Sag laccolith, Montana, U.S.A.

sediments. These are fed from subsurface feeders and no extrusive vents may occur. A submarine pile of pillow lavas and albite dolerites may thus be formed. Each individual pillow is, in effect, a small laccolith, and in aggregate these form larger laccoliths, which Jones and Pugh (1949) described as *laccolithic series*.

JOHN BRADLEY

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Cross-references: *Emplacement—modes; Hypabyssal rocks; Igneous intrusions—structural behavior.*

LAMPROITE

Lamproite is a group name for K- and Mg-rich lamprophyric volcanic rocks first used by Niggli (1923, see Tröger, 1935) to include the formerly described rocks *fortunite*, *verite*, and *jumillite* from Murcia Province, S Spain (Osann, 1906; Borley, 1967); *orendite*, *wyomingite*, and *madupite* from the Leucite Hills, Wyoming, U.S.A. (Cross, 1897); and *gaussbergite* from Antarctica (Lacroix, 1926; see Tröger, 1935, p. 322). Subsequently, another suite of these rocks (*fitzroyite*, *cedricite*, *mamilite*, and *wolgidite*)

was discovered in the W Kimberley district of W Australia (Wade and Prider, 1940; Prider, 1960).

Petrographically, lamproites range from coarse holocrystalline to vitrophyric (Prider, 1982), and the essential minerals present are various assemblages of leucite (or its altered equivalent—see Prider and Cole, 1942), sanidine, olivine, phlogopite, diopside, and magnophorite. Most are characterized by the presence of leucite, even though they are silica-saturated rocks containing normative quartz. The mineral composition of members of this group is given in Table 1. Accessory minerals in most of the lamproites include apatite, perovskite and priderite and rare accessories are wadeite and noonkanbahite (Prider, 1939, 1965).

Fuster et al. (1967) consider that jumillite, fortunite, and verite in S Spain are better distinguished on a chemical basis as jumillitic (with normative olivine and leucite), cancalitic (with normative olivine), and fortunitic (silica-saturated) lamproites, than on a mineralogical basis. For the field geologist, however, a mineralogical and textural classification is more applicable than a geochemical one, and so the multiplicity of names cannot be avoided. Likewise, while leucite–phlogopite lamproite (fitzroyite), leucite–diopside lamproite (cedricite) and leucite–magnophorite lamproite (mamilite) have essentially the same chemical composition, their mineral contents differ. So, if the cumbersome terminology such as leucite–olivine–phlogopite–diopside–magnophorite lamproite (= wolgidite) is to be avoided, the names given in Table 1 need to be retained.

Chemically, the lamproites are characterized by their K₂O proportion being in molecular excess over Al₂O₃, the dominance of K₂O over Na₂O (which is present only in very small

TABLE 1. Mineralogy of lamproites

Name	Essential minerals	Locality ^a
Cedricite	Leucite, diopside	WK
Fortunite	Sanidine, diopside–bronzite, phlogopite	MS
Fitzroyite	Leucite, phlogopite	WK
Gaussbergite	Leucite, diopside, phlogopite	Ga
Jumillite	Leucite, sanidine, olivine, diopside, phlogopite	MS
Madupite	Phlogopite, diopside	LH
Mamilite	Leucite, magnophorite	WK
Orendite	Leucite, sanidine, phlogopite, diopside	LH
Verite	Olivine, phlogopite, diopside	LH, MS
Wolgidite	Leucite, olivine, phlogopite, diopside, magnophorite	WK
Wyomingite	Leucite, phlogopite, diopside	LH, WK

^a GA, Gaussberg, Antarctica; LH, Leucite Hills, Wyoming, U.S.A.; MS, Murcia Province, Spain; WK, W Kimberley district, W Australia.

proportions in most of these rocks), and high Ti, Zr and Ba. Mineralogically lamproites are characterized by the occurrence of minerals such as magnophorite (a K–Mg amphibole), wadeite (K–Zr silicate) and priderite (K-titanate), which were first found in the W Australian leucite lamproites (Prider, 1939; Norrish, 1951). These minerals, which are not found in other rocks, have been recognized in all leucite lamproite suites (e.g., Carmichael, 1967). Other notable mineralogical features of these Ti-rich rocks (up to 7.3% TiO₂) include the absence of Fe–Ti oxides, which is due to the very high K (up to 12.6%) in the parent magmas.

The high-K, high-Mg ultramafic nature of lamproites gives these rocks an affinity with kimberlite, made more apparent by the discovery of a suite of olivine-bearing lamproites with significant diamond contents in the W Kimberley area of W Australia (Jaques et al., 1984). Rather than representing an intermediate group between lamproite and kimberlite, Scott Smith and Skinner (1984) consider the presence of leucite, Ti-rich, Al-depleted phlogopite, Ti-rich potassic amphibole, and sanidine indicates a true lamproite kinship for these rocks (diamondiferous lamproites). The type locality for diamondiferous lamproites is W Kimberley (Australia), but similar diamondiferous and nondiamondiferous rocks are known from Pipe Creek, Arkansas, U.S.A. and Holsteinborg, W. Greenland.

Present evidence points to lamproites having been derived by minor partial melting at depths of >100 km (Gupta and Yagi, 1980; Bergman 1986). The mantle peridotite of the source region shows geochemical characteristics indicative of being metasomatically enriched in K, Ba, REE, and related elements because of mantle “plumes” coming possibly from as deep as the core–mantle boundary (Arima and Edgar, 1983; Edgar, 1983). With the evidence that some lamproites are diamondiferous, these rocks are of deeper origin than any terrestrial rocks, except kimberlites, and it is likely that their genesis is largely or entirely independent of lithospheric plate motions.

R. T. PRIDER

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- Cross-references: *Alkaline rocks—undersaturated; Kimberlite; Lamprophyre; Leucitite; Ultramafic rocks.*

LAMPROPHYRE

Dark, lustrous rocks with mafic phenocrysts occurring in the Fichtelgebirge, W Germany, were called *lamprophyre* (shining porphyries) by C. W. von Gümbel. A few pre-existing names, some still in use, were incorporated by him in the new term. Later, Rosenbusch (1887) included it in his systematic classification as a group name for rocks with readily recognizable field characteristics, forming minor intrusions, usually dykes, associated with diorites and syenites. Owing to their unusual mineralogy and chemistry, they have given rise to a large number of rock names, most of which are now regarded as unnecessary. The more recent attempts to improve and simplify nomenclature include those of Wimmenauer (1973), Streckeisen (1979) and Rock (1979, 1984).

The following are general characteristics of the commoner lamprophyres, about which there is a large measure of agreement (Streckeisen, 1979).

- 1. Mineral composition and panidiomorphic texture separate them from the common plutonic and volcanic rocks sufficiently to justify a separate classification.
- 2. They are dark-colored rocks containing essential biotite (or Fe-rich phlogopite) and/or amphibole, with clinopyroxene and olivine, and in the cases of alnöite and polzenite, melilite.
- 3. Feldspars and feldspathoids, when present, are restricted to the groundmass.
- 4. Hydrothermal alteration is common.

- 5. Calcite, zeolites and other hydrothermal minerals may be primary constituents.
- 6. Contents of K₂O and Na₂O are comparatively high at medium to low SiO₂ content (<40–45%; in alnöite 30–35%).
- 7. Contents of H₂O, CO₂, S, P₂O₅, Ba, Sr, and some trace elements are high compared with other rocks of similar composition.

The IUGS scheme of classification distinguishes three main types of lamprophyre: (1) calc-alkaline (*minette*, *vogesite*, *kersantite*, *spessartite*), (2) alkaline (anchibasaltic of Wimmenauer, 1973) (*camptonite*, *sannaite*, *monchiquite*), and (3) melilitic (*alnöite*, *polzenite*) (Streckeisen, 1979). Mineral composition is summarized in Table 1.

Calc-alkaline lamprophyres

These are the lamprophyres *sensu stricto* of Wimmenauer (1973) and the shoshonitic lamprophyres of Joplin (1966) and Rock (1977). They are mainly associated with postorogenic granitoid complexes in orogenic belts. The typical calc-alkaline lamprophyres have color index >35. There are varieties with the same texture and the same minerals present but with a color index of <35 (commonly 25–35); these are the semilamprophyres of D’Amico (1961, 1962).

Minette consists of alkali feldspar (K and Na), biotite, diopsidic augite, plagioclase, hornblende, olivine, quartz, apatite, opaque minerals, calcite, and zeolites.

TABLE 1. Mineral composition of lamprophyres

	Olivine	Diopsidic augite	Titanaugite	Hornblende (green–brown)	Barkevikite– kaersutite	Biotite	Alkali feldspar	Albite–oligoclase	Andesine–labradorite	Feldspathoid	Glass	Melilite
Minette	*	*				**	**	*				
Vogesite	*	*		**			**	*				
Kersantite	*	*				**			**			
Spessartite	*	*		**			*		**			
Camptonite	*		**		**	*	*		**	*		
Sannaite	*		**		**	*	**		*	*		
Monchiquite	*		**		**					*	*	
Fourchite			**		**	*				*		
Alnöite	**	*				**						**

** essential; * usually present.
Accessory minerals excluded.

Vogesite consists of alkali feldspar (K and Na), hornblende (green and brown), diopsidic augite, plagioclase, biotite, olivine, quartz, apatite, opaque minerals, calcite, and zeolites.

Kersantite consists of plagioclase (oligoclase or andesine), biotite, augite, alkali feldspar, hornblende, olivine, quartz, apatite, opaque minerals, calcite, and zeolites; kersantite and vogesite are heteromorphous to a high degree.

Spessartite consists of plagioclase (andesine), hornblende (green and brown), diopsidic augite, alkali feldspar, biotite, olivine, quartz, apatite, opaque minerals, calcite, and zeolites.

Alkaline lamprophyres

These are the anchibasaltic lamprophyres of Wimmenauer (1973). Their chemistry is that of alkali basalts and basanites, and for some of them, of nephelinites, with the addition of volatiles. They are mainly associated with alkaline complexes, e.g., nepheline syenites and carbonatites. Their color index exceeds 40.

Camptonite consists of amphibole (barkevikite, kaersutite), titanaugite, olivine and/or biotite, in a groundmass of labradorite, amphibole, and pyroxene with subordinate alkali feldspar and feldspathoid, $\pm$ apatite, opaque minerals, calcite, and zeolite.

Sannaite is similar to camptonite, but with alkali feldspar dominant over plagioclase.

Tannbuschite is also similar to camptonite but with nepheline and feldspars, the nepheline being the predominant felsic mineral.

Monchiquite consists of titanaugite, amphibole (barkevikite, kaersutite), biotite, and commonly olivine, in a colorless isotropic groundmass of glass (potential plagioclase + nepheline) or analcime or nepheline, rich in microlites of pyroxene, amphibole, opaque minerals, apatite, calcite, and zeolites. A number of varieties are distinguished as follows: glassy base = hyalo-monchiquite; analcime base = analcime monchiquite; nepheline base = nepheline monchiquite; *fourchite* = olivine-free monchiquite; *ouachitite* = melanocratic biotite-free monchiquite.

Melilitic lamprophyres

These are mainly associated with alkaline complexes and carbonatites. Their color index exceeds 70.

Alnöite consists of melilite, biotite, with or without clinopyroxene, olivine, calcite, apatite, perovskite, opaque minerals, and nepheline, and is feldspar-free. Color index is greater than 90.

Polzenite (including *modlibovite*, *vesecite*,

luhite, *bergalite*) consists of melilite, biotite, essential feldspathoid (usually nepheline or h  yne), with or without olivine, titanaugite, calcite, apatite, perovskite, and opaque minerals. It is feldspar free. Color index is 70–90.

Some authors recognize a fourth group of lamprophyres, the *lamproites*, which are not mentioned directly in the IUGS scheme. They are dark-colored rocks rich in K and Mg, texturally resembling lamprophyres but with some members containing phenocrysts of feldspathoid minerals, especially leucite (see **Lamproite**).

Occurrence

Lamprophyres occur most commonly as narrow dykes, sills, sheets, and plugs, and more rarely as lavas. Close association with igneous centers in which granites, diorites, and syenites, both feldspathoidal and feldspathoid-free, is common but not universal. Evidence of explosive emplacement may be present. The typical texture is panidiomorphic with mafic minerals occurring both as phenocrysts and in the groundmass. Ocellar and globular structures are common (see Rock, 1984, table 6). Xenocrysts, sometimes of large size, of alkali feldspar, amphibole, pyroxenes, biotite, and apatite have been widely recorded, as have xenoliths of mantle or deep crustal provenance. Where they have been emplaced before the cessation of earth movements, intrusions of lamprophyre are more readily sheared than those of the country rock, giving rise to "lamproschists."

Chemical composition

Chemical analyses of eight of the commoner types of lamprophyre are given in Table 2. It should be noted that while the available data suffice to distinguish monchiquites and camptonites from other lamprophyres and possibly from each other, in terms of average chemical composition, minettes, vogesites, kersantites, and spessartites seem virtually indistinguishable.

For the alkaline lamprophyres (camptonite, sannaite, tannbuschite and monchiquite), Rock (1977) has given new definitions and proposed a chemical screen for all alkali lamprophyres. He has also reviewed previous definitions. Rock (1984) proposes detailed modal, mineralogical, chemical, and normative screens for the calc-alkaline lamprophyres.

Petrogenesis

Studies of lamprophyre petrogenesis have indicated a very wide range of possible

TABLE 2. Composition of common lamprophyres; number of analyses in parentheses.

	1 (140)	2 (31)	3 (179)	4 (129)	5 (95)	6 (90)	7 (9)	8 (6)
SiO ₂	51.36	50.70	51.45	51.94	42.96	41.06	42.53	31.58
TiO ₂	1.46	1.44	1.16	1.36	2.75	2.63	3.14	2.54
Al ₂ O ₃	12.65	14.01	14.77	15.12	14.82	13.20	13.87	8.88
Fe ₂ O ₃	3.04	3.36	2.62	2.93	4.64	4.39	5.50	5.12
FeO	4.10	5.04	5.06	5.46	7.44	7.16	6.19	6.16
MnO	0.14	0.17	0.12	0.14	—	—	—	—
MgO	7.27	7.41	6.24	6.85	6.68	8.88	5.01	18.67
CaO	6.37	6.83	5.99	7.04	10.18	11.15	10.88	17.64
Na ₂ O	1.91	2.76	2.95	3.19	3.41	3.25	3.75	1.30
K ₂ O	6.08	4.19	3.36	2.40	2.02	1.99	2.23	2.12
P ₂ O ₅	1.08	0.77	0.66	0.44	0.71	0.84	0.86	1.51
H ₂ O	2.12	2.39	2.76	2.40	—	—	—	1.68
CO ₂	2.23	1.79	1.74	1.10	2.38	2.13	1.20	2.51
TOTAL	99.81	100.86	98.88	100.37	97.99	96.68	95.16	99.93

1, minette; 2, vogesite; 3, kersantite; 4, spessartite; 5, camptonite; 6, monchiquite; 7, fourchite; 8, alnöite.
1–4 from Rock (1984), 5–7 from Rock (1977), 8 from Nockolds et al. (1978).

controlling factors, including a variable but significant crustal contribution to partial melts from the mantle. The depth at which melting takes place, the nature of the mantle that was melted and the proportion of melt vs. residue are matters of debate, as are the relative depletion or enrichment of particular elements, the CO₂/(CO₂ + H₂O) ratio of the source region and the degree to which fractional crystallization was operative. However while the origin of calc-alkaline lamprophyres, for example, has been referred to as “obscure and contentious” (Rock et al., 1986), it seems clear that a satisfactory explanation of petrogenesis of any particularly lamprophyre suite will involve the interplay of a number of factors (e.g., Macdonald et al., 1985). A review of the existing state of knowledge is presented by Rock (1987).

Other terms

- Alvikite*—a quartz-bearing melanocratic lamprophyre rich in calcite and biotite, sometimes with nepheline in the matrix (Alvik, Sweden).
- Autosite*—compositionally similar to kersantite but without feldspar.
- Bermudite*—a lamprophyric volcanic rock with small biotite phenocrysts in an analcime- and nepheline-bearing matrix; effusive equivalent of biotite monchiquite or ouachitite.
- Calcikersantite*—a kersantite containing labradorite or bytownite.
- Camptospessartite*—a dark-colored spessartite with titanaugite as the pyroxene.

- Cascadite*—a minette with phenocrysts of biotite, olivine, and augite in a groundmass mainly composed of alkali feldspar.
- Cuselite*—an augite-biotite-bearing lamprophyre with a composition between that of minette and vogesite and between kersantite and spessartite.
- Epichellite*—similar to camptonite with phenocrysts of hornblende, augite, olivine, magnetite, and pyrite in a groundmass that may contain analcime; It resembles teschenite but has less analcime and is porphyritic (Epichel, Portugal).
- Eustratite*—a lamprophyric rock with few phenocrysts of olivine, augite, and corroded hornblende in a magnetite-interstitial feldspar-mica-glass groundmass.
- Garéwaite*—nearly feldspar-free, composed of phenocrysts of diopside in a holocrystalline groundmass of olivine, pyroxene, chromite, and magnetite; a peridotite porphyry related to vogesite-spessartite-odenite.
- Garganite*—a vogesite containing both augite and hornblende.
- Giumarrite*—an amphibole-bearing monchiquite (Giumarra, Sicily).
- Gladkaite*—see **Aplite**.
- Hamrongite*—a dark-gray, fine-grained quartz kersantite with phenocrysts of black mica in a groundmass with intersertal texture and containing mica, andesine and some quartz.
- Heumite*—a mesocratic nepheline-bearing lamprophyre with barkevikite, biotite, alkali feldspar and sodalite.
- Holmite*—similar to monchiquite but with a groundmass of melilite rather than analcime.

Kvellite—a dark-colored ultramafic lamprophyric rock with phenocrysts of lepidomelane, olivine, barkevikite, apatite, ilmenite, and magnetite.

Lamproschist—metamorphosed lamprophyre with schistose texture.

Malchite—rare phenocrysts of hornblende, labradorite, and sometimes biotite in a groundmass of hornblende, andesine, and minor quartz (Malchen, E Germany).

Minette-felsite—a minette-like rock but with a microcrystalline or cryptocrystalline rather than a phanerocrystalline groundmass.

Mondhaldeite—similar in composition to camptonite with phenocryst of acicular hornblende, augite, bytownite, and leucite in a glassy groundmass (Mondhalde, W Germany).

Odinite—a greenish-gray rock with phenocrysts of labradorite, augite (or diallage), and sometimes hornblende in a groundmass of lath-shaped or equigranular feldspar and a felty mesh of acicular hornblende crystals.

Orbite—phenocrysts of hornblende (sometimes with plagioclase) in a groundmass with abundant plagioclase and having the composition of malchite (Orbishohe, W Germany).

Raabsite—a minette containing sodic amphibole (Raabs, Austria).

Tjosite—dark-colored, porphyritic nepheline-bearing and intermediate in composition between a lamprophyre and a syenite (Tjose, Norway).

Topsailite—phenocrysts of plagioclase (ca. An₅₀) augite, apatite, and titanoferrite, in a groundmass composed of andesine, biotite, barkevikite, augite, and sphene.

Turjite—a melanocratic lamprophyre rich in analcime, biotite, calcite and melanite; an analcime-calcite alnöite (Turij Mis (Turja) Kola, U.S.S.R.).

Volhynite—a quartz-bearing kersantite composed of phenocrysts of plagioclase, hornblende, and sometimes biotite in a groundmass of quartz, feldspar, and abundant chlorite (Volhynia, U.S.S.R.).

A. HERRIOT

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Cross-references: *Alkaline rocks—undersaturated; Appinite; Basalt; Breccia pipe; Calc-alkaline rocks; Carbonatite; Hypabyssal rocks; Kimberlite; Lamproite; Melilitite; Nephelinite; Textures of igneous rocks.*

LATITE

Latite is a porphyritic effusive rock composed of phenocrysts of plagioclase and K-feldspar (sometimes sanidine) in nearly equal proportions, little or no quartz, and a finely crystalline or glassy groundmass. Clinopyroxene is the most common ferromagnesian mineral, but orthopyroxene, olivine and, less commonly, hornblende may occur, while biotite is commonly present. Where crystalline, the groundmass may be trachytic with laths of sanidine and plagioclase, Fe-ore, and some mafic constituents, commonly clinopyroxene. A quartz latite is a *rhyodacite*.

The term latite is used in the classification of common volcanic rocks by Streckeisen (1979) for rocks between trachyte and andesite-basalt with many authors considering latites to be equivalent to trachyandesites and trachybasalts. However, some authors assign latites to the

shosonitic rock series, with relatively high K_2O and a chemical composition between andesite and trachyandesite (Middlemost, 1985). Other authors distinguish latites (volcanic equivalents of monzonites) and latite-andesites (volcanic equivalents of monzodiorites) from lavas corresponding with alkali monzonites, monzodiorites, or monzogabbros (K-rich series of tristanite, trachyandesite, and trachybasalt; more Na-rich series of benmoreite, mugearite, and hawaiite), i.e., distinguish latite from trachyandesite and trachybasalt (Nockolds et al., 1978).

D. R. BOWES

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Cross-references: *Andesite and dacite*; *Basalt*; *Igneous rocks—classification and nomenclature*; *Intermediate igneous rocks*; *Monzonite (syenodiorite)*; *Trachyandesite, trachybasalt*; *Trachyte*; *Volcanic rocks*.

LAVA AND LAVA ERUPTIONS

Lava

Lava is the form in which magma is erupted onto the Earth's surface. It consists of three kinds of material: a silicate melt, suspended solids, and gases. The gases are usually present as bubbles, for under surface pressure lava can dissolve little volatile material; the bubbles represent mainly the gases that were dissolved in the magma at depth. The solids are mostly crystals of minerals that have crystallized from the melt, rock fragments picked up from the surface or from the walls of the conduit, and engulfed fragments of the surficial crust of the lava flow. The essential constituent is the silicate melt, which may form up to 100% of the lava; its relative abundance determines the ability of the lava to move after its eruption. This melt decreases in amount following eruption, as falling temperature brings about its solidification, in many cases to a glass rather than to crystalline material.

The viscosity of a lava, probably its most important physical property, controls to an important degree the rate of flow of the lava,

the thickness of the flow, the surface features of the flow, and the amount of glass in the resultant rock. Two different viscosities are involved: the viscosity of the melt, which controls the amount of glass formed, and the bulk viscosity of the lava, which controls the other three features. The viscosity of the melt depends mainly on its composition, for under surface conditions it should contain little dissolved volatile material. In general, the viscosity increases with the silica content of the melt. The effect of temperature cannot be separated from that of composition, for the melts that exist at low temperatures are those that are rich in silica. For a given melt, however, the increase in viscosity during solidification may be very great. The viscosity of the lava as a whole depends in part on that of the melt present, but it also depends on the relative abundance of solid material and of gas bubbles: both tend to increase the viscosity. The viscosity of an olivine basalt melt is about 3,000 poise (viscosity of water = 0.01 poise); that of a partly crystalline vesicular lava must be somewhat greater.

Lava eruptions

The eruption of a contiguous mass of lava is usually accompanied by some pyroclastic eruptive activity. However, the outpouring of lava can usually be distinguished from the more explosive activity and can be treated separately. Distinguishing first between subaerial and subaqueous eruptions, it is possible to further classify the resultant flows on their surface structure and thickness, both features that depend largely on the viscosity of the lava. On the basis of this viscosity, these eruptions can be grouped into effusions and extrusions.

Effusions. Subaerial effusion of the most fluid lavas produces *pahoehoe* flows (*ropy lava*), characterized by relatively smooth to billowy surfaces. Such flows are generally of the order of 3 m thick. Locally, areas of small-scale wrinkling may occur on the surface ("festooned pahoehoe"), but such structure is not widespread. Other patterns have resulted in such terms as corded pahoehoe, shelly pahoehoe, slab pahoehoe, etc. In many cases a marginal zone composed of tilted slabs of the flow crust apparently develops at a late period by lateral pressure from the central parts of the flow. Movement of the lava to the front of the flow is largely through tubes within the flow, so that relatively little heat is lost until the flow front is reached. Here the flow usually advances by the protrusion of successive lobes or "toes" of lava that quickly develop an elastic glassy skin. The lobes expand as lava flows into them from

behind until the skin ruptures, allowing a new lobe to be protruded. On steep slopes the flow may advance with a rolling motion, the skin being carried forward, down, and under the flow by its advance. The very fluid nature of these flows restricts them to basaltic lavas.

Effusion of somewhat more viscous lava results in an *aa* flow (*blocky lava*, *apalhraun*), usually of basaltic to andesitic composition. Such flows are usually thicker than pahoehoe flows—7–8 m on the average. The surface of an *aa* flow is characteristically a jumble of angular blocks (“clinker”) covered with short spines. A section through such a flow would show that the clinker forms only the upper and lower layers, separated by a layer of compact rock. Moreover, most of the clinker below the surface is firmly welded together and to the compact layer. The lava is carried to the flow front by an open river bordered on either side and at the front by a “moraine” of clinker, actually a sheet of clinker resting on a sheet of still-mobile material (Fig. 1). The clinker itself results from the fracturing of the thick crust that develops; as the blocks are pulled apart by movement of the underlying liquid, strings of plastic lava stretch between them and finally break—hence the spines. The advance of the flow results in the gradual steepening of the clinker-covered front until it becomes unstable and masses of clinker slide off, revealing for a moment the incandescent plastic core of the flow. This quickly cools, darkens, and breaks up into a new veneer of clinker, as the flow slowly advances over the clinker deposited in front of it, the front steepening so as to repeat the process. A similar process goes on along the sides of the flow, though lateral spreading is not so rapid as the advance.

Extrusions. Extrusions differ from effusions only in the greater viscosity of the lava in extrusions; this in turn causes extrusions to spread less widely and to form thicker masses with steeper fronts and sides and it causes the rocks of which they are composed to contain a greater proportion of glass or cryptocrystalline material. It also affects the character of the surface of these lava masses: they are covered by a rather thick veneer of angular blocks, the individual blocks being smooth-sided rather than spiny, leading to their characterization as block (blocky) lava.

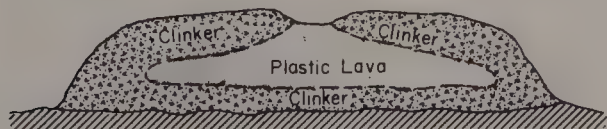


FIGURE 1. Section through an *aa* flow; the layer of plastic lava eventually solidifies to form the compact zone in the flow.

The less viscous extrusions usually occur in the form of coulees or of cow-pies (or cow-pats). *Block-lava coulees* develop where such extrusions occur on a sloping land surface (such as the flank of a volcano), which controls the movement of the lava and leads to the production of a tongue-like flow. *Block-lava cow-pies*, on the other hand, develop where extrusion is onto a relatively flat surface (such as the floor of a caldera) and the lava spreads (albeit slowly) more or less equally in all directions to yield a cake-like mass. Both coulees and cow-pies reach thicknesses of 30 m or more and may stretch for several kilometers from the vent. They usually are of andesitic or dacitic composition. Movement is similar to that in *aa* effusions. Large spines may develop on the upper surface of the flow, these pinnacles of plastic lava apparently being squeezed from between the blocks by the pressure of the lava.

The more viscous extrusions characteristically produce *extrusion (endogenous) domes*. These steep-sided, more or less circular mounds, with an average height of 50–150 m, are usually composed of glass-rich dacite, rhyodacite, and rhyolite. The viscosity of the lava extruded is too great to allow flowage away from the vent, and the lava simply piles up around it. Lava is added to the mass from below, gradually inflating it and breaking the rigid crust into smooth angular blocks. In this way a thick mantle of talus is produced that conceals the compact core of the dome. Occasionally portions of the dome may break away from its steep flanks and roll to the base as an incandescent avalanche. The fragments produced may still be plastic enough to vesiculate and explode, releasing quantities of gas that may increase the mobility of the avalanche and convert it into an *agglomerate flow* in which each particle is surrounded by an envelope of hot expanding gases that isolates it from its neighbors and so reduces the internal friction of the mass almost to zero. Spines are even more characteristic of extrusion domes than of cow-pies and coulees.

Subaqueous eruptions of very fluid (basaltic) lava generally produce flows of *pillow lava*. These develop instead of pahoehoe under conditions of rapid chilling and a steep flow front. The rapid chilling develops a strong glassy skin on the lobes, which tend to break away from the rest of the flow as a result of the steep slope on which they lie. There thus forms a pile of bladder-like masses of lava that accommodate their shapes to one another before their interiors have solidified. The spaces between the pillows may be filled with a variety of materials particularly mud, glass fragments and diatomaceous ooze, etc. It is possible for pillow lavas to

develop even under certain subaerial conditions.

If the lava is somewhat less fluid, the tendency of the crust to break up with movement of the interior liquid gives water access to this hot liquid, resulting in violent steam explosions. Thus, the lava is quickly chilled and simultaneously blown into small fragments, producing a deposit called *hyaloclastite* consisting of various-sized fragments of glass in a matrix of very fine-grained clayey material.

Other terms

Blister—a superficial swelling up of the crust of a lava flow resulting from puffing up of gas or vapor.

Blow-hole—a small crater on the surface of a viscous lava flow.

Epimagma—a semisolid, relatively gas-free, vesicular, magmatic residue commonly formed by the cooling of lava in a lava lake.

Hydroexplosion—an explosion resulting from the contact of a molten lava flow with water.

Lava tube—a tubular-shaped mass of solidified lava formed by withdrawal of molten lava after the formation of the superficial crust.

Lekolith—an effusive rock mass formed as the result of congealing of a lava lake (Greek *lekos* = dish and *lithos* = stone).

Peperite—either the result of mixing of lava and marine sediment or of intrusion of magma into wet sediment; the resultant rock is breccia-like.

Stalactite—a cone-shaped mass of lava hanging from the roof or walls of a lava tunnel or other cavity and the result of dripping of fluid lava.

Stalagmite—a cone-shaped mass of lava built up from the floor of a cavity in a lava flow; a corresponding feature to a stalactite of lava.

Tongue—a lava flow that may be as much as several kilometers in length and is an offshoot from a larger flow.

Tufflava (tuffolava)—an extrusive rock intermediate between a lava flow and a welded-tuff type of ignimbrite.

Tumulus—a dome or small mount on the crust of a lava flow; unlike a blister, it is a solid structure.

CHARLES P. THORNTON

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- Cross-references: *Endogeneous dome*; *Explosive volcanic eruptions—classification*; *Magma*; *Volcanic glass*; *Volcanoes and volcanism*.

LAWSONITE-ALBITE-CHLORITE FACIES

The very low-grade metamorphic lawsonite-albite-chlorite facies is distinguished from the glaucophane schist (blueschist) facies and the prehnite-pumpellyite facies by the presence of the high-density hydroxylated Ca-Al silicate, lawsonite, in coexistence with albite and chlorite and in the absence of glaucophanic amphibole or jadeitic pyroxene. Aragonite and/or calcite, epidote, pumpellyite, and in some cases prehnite, may also be present as well as phengitic muscovite, titanite (sphene), and other minerals. Laumontite-bearing assemblages of the zeolite facies pass directly into those of the lawsonite-albite-chlorite facies by the reaction



This reaction occurs at a pressure of about 3 kb in the general temperature range ca. 200–250°C with comparatively little temperature dependence. The facies is one of low temperature, and relatively high pressure:temperature ratio. Pelitic, metagraywacke, and some calcareous assemblages of the facies are developed on a regional scale in New Zealand, northern California, U.S.A., New Caledonia, and elsewhere (Turner, 1981). The newly-formed minerals in the rocks concerned are commonly very fine-grained and may define a more or less well-defined cleavage.

Winkler (1965) used the term *lawsonite-albite facies* to include assemblages with Na-amphibole, as well as those without. The lawsonite-albite-chlorite facies represents a lower grade part of Winkler's facies, in which albite and chlorite coexist without reaction to glaucophanic amphibole.

DOUGLAS S. COOMBS

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Cross-references: *Glaucophane schist facies*; *Metamorphic facies*; *Regional metamorphism*; *Zeolite facies*.

LEUCITITE

Leucitite is an undersaturated, extrusive or hypabyssal, fine-grained or porphyritic rock composed essentially of leucite and pyroxene first defined by H. Rosenbusch in 1877 (see Tröger, 1935). Accessories include nepheline, melilite, olivine, apatite, iron ores, and (rarely) plagioclase. When olivine is present in amounts exceeding 5%, these rocks are referred to as leucite basalt, but would be better described as olivine leucitite. Nepheline is common and all transitions exist between leucitite and nephelinite.

The name leucitite would imply a monomineralic rock composed entirely of leucite but, like the names nephelinite and melilitite, it was first applied to pyroxene-bearing rocks. Indeed, when a rock consisting almost wholly of leucite was discovered, it was necessary to coin the name *italite* (Washington, 1920) for it because the most suitable name (leucitite) was already taken. In order to overcome the inconsistencies in nomenclature of the feldspar-free lavas, Hatch et al. (1972) suggest that the name leucitite be replaced by the name *leumafite*.

Rocks bearing close similarities to olivine leucitites are (1) *ugandite*, a melanocratic olivine leucitite associated with the rare kalsilite-bearing *katungite* lavas of Uganda (Holmes, 1937), and (2) *missourite* and *fergusite* from Highwood Mountains, Montana, U.S.A. (Tröger, 1935), which are coarser-grained, of deeper-seated crystallization, and with some of the leucite replaced by orthoclase yielding pseudoleucite.

Leucitites are of small bulk compared with other volcanic rocks. The best-known localities are the Roman Province of W Italy and the Birunga and Toro-Ankole volcanic fields of Uganda. An average analysis (Nockolds, 1954) is SiO₂ 47.11, TiO₂ 1.25, Al₂O₃ 15.74, Fe₂O₃ 4.54, FeO 4.54, MgO 5.24, CaO 11.01, Na₂O 2.02, K₂O 6.72, P₂O₅ 0.44, H₂O+ 0.87.

Sheraton and Cundari (1980) refer to the lamproite (gaussbergite) of Gaussberg in Antarctica as leucitite. Their work on the Gaussberg volcanic rocks shows the close relation between leucitites and lamproites.

The petrology of the leucitites and other leucite-bearing rocks is described by Sørensen (1974) and Gupta and Yagi (1980).

The following are related rock types.

Albanite—a leucitite containing phlogopite.

Batukite—an olivine melaleucitite.

Braccianite—a variety of leucitite with leucite, augite, nepheline, olivine, anorthite, magnetite and apatite (Bracciano, Italy).

Cecilite—similar to braccianite but with abundant melilite.

Etindite—a dark-colored effusive rock intermediate in composition between leucitite and nephelinite.

Fiasconite—an anorthite-bearing leucitite-basanite that also contains olivine, augite, nepheline, and Fe-ores (Montefiascone, Italy).

Leucitolith—an effusive rock composed almost entirely of leucite.

Leucitophyre—a porphyritic effusive rock mainly composed of leucite, nepheline and augite.

Mafurite—a variety of olivine leucitite in which kalsilite is present instead of leucite.

Mikenite—a leucitite having higher Na₂O/K₂O than ordinary leucitite.

Nemite—a melaleucitite (Lake Nemi, Italy).

Sperone—a vesicular leucitite containing small melanite crystals.

Tusculite—a melilite leucitite with small proportions of pyroxene, ilmenite and feldspar (Tusculum, Italy).

Ugandite—an olivine leucitite with a glassy base (Bufumbira, Uganda).

Vesbite—a extrusive rock intermediate in composition between leucitite and melilitite, being composed of leucite, melilite, acmite-augite and accessory apatite and opaque oxides.

R. T. PRIDER

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Cross-references: *Alkaline rocks—undersaturated; Basalt; Feldspathoidal rocks—genesis; Hypabyssal rocks; Peralkaline igneous rocks; Rift valley volcanism.*

LHERZOLITE

Lherzolite is an ultramafic rock first identified at the Etang du Lherz in the Pyrenees in 1797, and first described from this and other Pyrenean localities by Lacroix (1894). The original rock consists of olivine, two pyroxenes—a calcic clinopyroxene and a magnesian orthopyroxene—and accessory spinel. According to the IUGS recommended nomenclature (Streckeisen, 1976), lherzolite is a peridotite (40% olivine) with two pyroxenes (one ortho-, one clino-). The presence of various accessory phases like spinel, garnet, anorthite, amphibole, or phlogopite permits the definition of subspecies whose names are qualified by the appropriate mineral suffix(es), e.g., garnet lherzolite, spinel lherzolite.

Mineral composition may vary considerably, but as a rule the olivine and orthopyroxene are magnesian, and the garnet is rich in the pyrope molecule. The spinel is a member of the hercynite–spinel–chromite series. The clinopyroxene is usually diopsidic, and occasionally Cr-rich. Both it, the orthopyroxene and any amphibole present, are highly aluminous.

When garnet is present it is often surrounded by a kelyphitic reaction rim consisting of a complex intergrowth that may include pyroxene, garnet, amphibole, anorthite, and spinel. This variation reflects the sensitivity of this assemblage to changes in pressure and temperature. Experimental work has permitted the definition of a range of P – T facies that relate to upper mantle conditions (O'Hara, 1967). As a sample of mantle material, lherzolite approximates in composition to pyrolite—the hypothetical starting material for the genesis of basic and ultramafic magmas—and is considered to represent “fertile,” or nearly pristine mantle material (Morse, 1980).

Lherzolite occurs in three principal settings: (1) as part of the ultramafic lower portion of ophiolite complexes, (2) as xenoliths in kimberlite pipes, and (3) as inclusions in alkaline basalts (Harte, 1983; Menzies, 1983). In all three settings all of the P – T facies mineral assemblages have been identified. This appears to reflect the presence of a lherzolite component common to both suboceanic and subcontinental mantle although isotopic and

trace-element geochemical studies indicate some heterogeneity (Hawkesworth et al., 1983; Menzies, 1983). The ubiquitous occurrence of a tectonic fabric suggests that this layer has also undergone plastic deformation (Nicolas et al., 1971).

Cumulate material in some layered basic intrusions may have a lherzolite mineralogy (e.g., the Dawros intrusion, Eire–Rothstein, 1957), but massive bodies on the continents are a part of ophiolite complexes. Their margins are usually tectonic, but the presence of narrow, intense, high-temperature thermal aureoles often indicates hot emplacement. Within the lherzolite sole may be incorporated mineral assemblages of an eclogitic character reflecting decoupling of the ultramafic material from the upper mantle at considerable depth (e.g., Treloar et al., 1980).

ADRIAN F. PARK

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Cross-references: *Aureole*; *Basic igneous rocks*; *Eclogite*; *Flood basalt*; *Harzburgite*; *Kimberlite*; *Mafic and ultramafic inclusions in igneous rocks*; *Mid-ocean ridge and ocean-floor petrology*; *Peridotite*; *Plutonic rocks—classification and nomenclature*; *Pyroxenite*; *Reaction rim*; *Ultramafic rocks*; *Upper mantle—experimental petrology*.

LIMBURGITE

Limburgite is porphyritic volcanic rock consisting of phenocrysts of olivine, titanite, and opaque iron oxides in an alkali-rich glassy groundmass from Limburg, Kaiserstuhl, W Germany, first defined by H. Rosenbusch in 1872. According to Tröger (1935) different varieties contain microlites of amphibole, haüyne, or leucite in the glassy groundmass. Garry et al. (1972) give clinopyroxene, olivine, and opaque oxides as microlites in an alkali-rich glassy groundmass with some nepheline and/or analcime as possible constituents, but feldspars are typically absent. However, re-examination of the type-rock has shown that only one phase of the mass is feldspar-free; other phases contain basic plagioclase in the groundmass and thus limburgite is an ultramafic basalt containing accessory or occult basic plagioclase. Hatch et al. in earlier editions included it in their group of “fine-grained ultramafites,” but in 1972 omit all reference to limburgite, as does Johannsen (1938). Holmes (1928) equates limburgite with “magma basalt” or (chemically) “nepheline basalt with glass in place of the nepheline.”

The term is considered to be useful if it is confined to lavas with an alkalic glassy nepheline-free groundmass. Amongst described localities are Birunga, Uganda (Holmes and Harwood, 1937), Macedon, Victoria, Australia (Edwards, 1938a), the Kerguelen Archipelago (Edwards, 1938b), Otago, New Zealand (Hutton, 1942), Monte Rizzoni, Italy where there is a local variety called *rizzonite* (see Tröger, 1935), and Limburg for which the following chemical analysis is quoted by Tröger (1935): SiO₂ 41.44, TiO₂ 3.36, Al₂O₃ 14.16, Fe₂O₃ 5.97, FeO 6.15, MgO 8.67, CaO 12.22, Na₂O 2.01, K₂O 2.77, P₂O₅ 0.31, H₂O+ 1.67.

R. T. PRIDER

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Cross-references: *Alkaline rocks—undersaturated*; *Basalt*; *Peralkaline igneous rocks*; *Rift valley volcanism*; *Ultramafic lavas*.

LIQUID IMMISCIBILITY

Liquid immiscibility is the coexistence of two or more liquid phases in equilibrium. Petrogenetic interest has stemmed from the potential of a homogeneous magmatic liquid to separate a second liquid phase in response to a change in temperature, pressure or composition. Immiscibility occurs when the sum of the free energies of two melts is less than that of a mixture of them; it should not be confused with the failure of some contemporaneous magmas to mix for kinetic reasons, as seen in certain composite bodies of acidic and basic rocks.

Although the immiscibility of silica-rich and silica-poor melts was observed in melting experiments on binary metal oxide–silica systems by J. W. Greig more than sixty years ago, it appears restricted to temperatures greater than 1650°C and to compositions with less than a few percent Al₂O₃, Na₂O, and K₂O. At around the same time, N. L. Bowen refuted contemporary models invoking immiscibility, arguing from basic principles that postulated examples did not show the textural and mineralogical features expected from equilibrium growth and that the most convincing evidence for immiscibility, the coexistence of quenched glasses of contrasting composition, had not been presented.

The revival of interest in immiscibility which

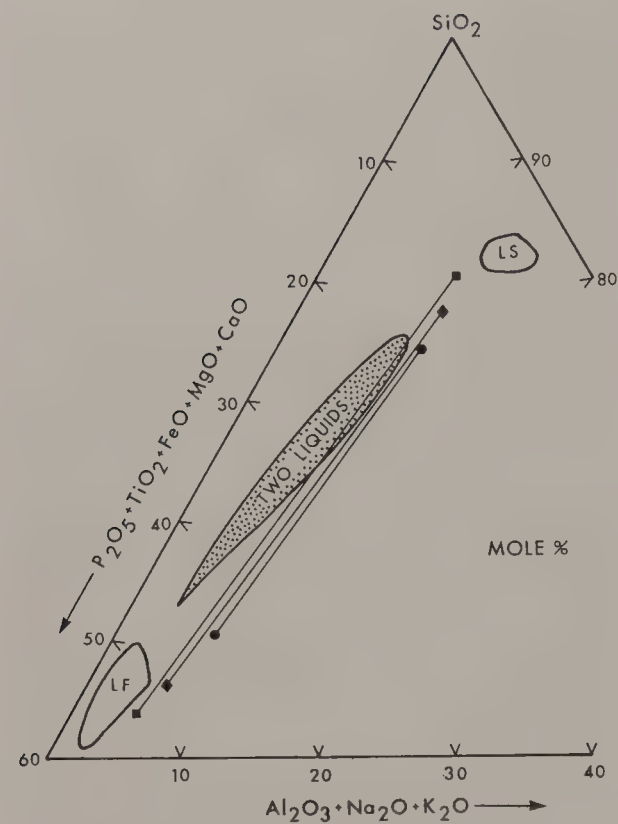


FIGURE 1. Compositions of immiscible glasses from the mesostases of basaltic rocks (LS and LF), compared with the immiscibility field in the join fayalite-leucite-silica (stippled) and compositions of immiscible melts produced by experimental fractionation of basaltic liquids (joined by tie-lines). (From data of various authors; after Freestone, 1979).

has taken place in the last twenty years is based upon developments which appear to answer many of the objections of Greig and Bowen. A two-liquid field at temperatures and in a compositional range more typical of magmas was delineated by E. Roedder in the system fayalite-leucite-silica. Furthermore, coexisting glasses of silica-rich (“granitic”) and iron-rich (“ferrobasaltic”) composition were widely recognized in the mesostases of tholeiitic basalts; they represent the extension of the immiscibility field discovered by Roedder into natural compositions (Fig. 1). The immiscibility has been extensively studied in the laboratory (summarized by Naslund, 1983) but its large scale petrogenetic significance is not fully resolved, although experimental work suggests that it may be responsible for certain granophyres of the Skaergaard intrusion. The miscibility gap is intersected in the late stages of crystallization of magmas undergoing a trend of pronounced Fe-enrichment, and is limited to a narrow range of temperature and oxygen partial pressure (Philpotts and Doyle, 1983). An additional difficulty is that many of the glassy droplets

observed in basalts are extremely fine, and may represent metastable growth in the supercooled liquid (Kuo et al., 1986), so that the apparent ubiquity of the phenomenon may be deceptive.

Attempts have been made to demonstrate that spheroidal textures in crystalline rocks meet the criteria for immiscibility, with varying success. In particular Philpotts (e.g., 1976) has argued that felsic “ocelli” in alkaline gabbroic dykes and sills of the Monteregian province, Quebec, Canada, are quenched immiscible droplets and that immiscibility may account for the association of syenite and gabbro in some cases. However, the crystalline nature of the ocelli permit alternative explanations such as the infilling of vesicles by residual liquids. Other postulated examples of immiscibility, for example in Precambrian volcanic rocks showing varying degrees of metamorphism, are in many cases tenuous and have not attracted widespread attention.

In addition to the immiscibility of two silicate liquids, miscibility gaps have been demonstrated in a large number of systems of some petrological interest where the anionic component of the second liquid is not a silicate but a sulfide, chloride, fluoride, phosphate, etc. For example, the immiscibility between transition metal sulfide liquids and silicates may be important in the genesis of economic Ni-Cu

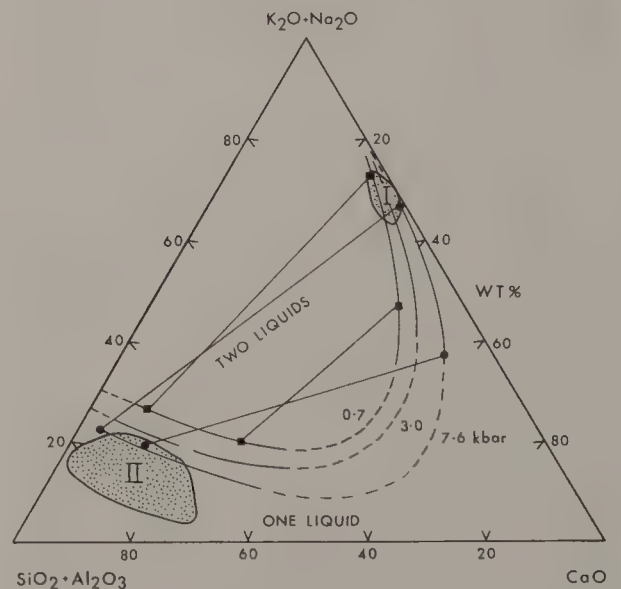


FIGURE 2. The outline of the two-liquid field produced by melting mixtures of phonolite, nephelinite and carbonates, at various pressures. Carbon dioxide and minor components such as MgO and FeO are neglected in the diagram. Selected conjugate immiscible melts are shown, joined by tie-lines. The stippled fields represent natrocarbonatite magma (field I) and alkaline silicate magmas including phonolites, nephelinites, alnöites, urtites etc. (field II). (After Freestone and Hamilton, 1980.)

deposits (Rajamani and Naldrett, 1978). A substantial two-liquid field exists between carbonate and silicate liquids, which shows a strong pressure dependence (Fig. 2), and it has been proposed that carbonatite magmas form as the result of immiscible separation from magmas of nephelinitic to phonolitic composition. The precise compositions of the silicate parent and the carbonatite derivative are a matter of some discussion, however (Twyman and Gittins, 1987).

IAN C. FREESTONE

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- Cross-references: *Alkaline rocks—undersaturated; Basalt; Carbonatite; Experimental petrology; Granophyre; Magma; Skaergaard intrusion; Textures of igneous rocks.*

LOPOLITH

A *lopolith* is a large, concordant, typically layered, igneous intrusion of planoconvex or lenticular shape that is sunken in its central part because of sagging of the underlying country rock (from the Greek *λοπας*, a basin or flat earthen dish). The term was originally proposed by Grout (1918) to describe the Duluth gabbro; the Sudbury and Bushveld intrusions were cited among other examples. Intrusions subsequently included in lists of lopoliths include the Great Dyke, the Insizwa, Freetown, and Kungwe Bay

intrusions in Africa, and the Stillwater intrusion in the U.S.A.

While the original concept was that of a concordant, basin-like structure, much thinner than it is wide, it has long been recognized that the lower contacts generally dip quite steeply, although irregularly, towards the center. The typical structure is thus funnel-shaped (Wilson, 1956): *ethmolith* would be a more accurate name. In fact “lopolith” already has an archaic ring, and it is noticeable that it is not used by workers in any of the “type” areas.

Apart from their external form and great size (the Bushveld Complex covers 65,000 km²), the most typical feature of structures described as lopoliths were produced by physical and chemical differentiation of the magma. The layers may range from ultramafic and mafic rocks—peridotite, pyroxenite, norite, gabbro—below, to more felsic ones—including granophyric “red-rocks”—at the top. Both rhythmic and cryptic layering may be found, on both large and small scales. The layering generally strikes parallel to the contacts but dips less steeply than the base (see *Astrobleme* for map).

External contacts are concordant in a gross way, but may be highly irregular in detail. Moreover, the concordance is mainly one of strike. Because of the enormous volumes of magma involved, metamorphic aureoles may be very wide. Brecciation, stoping, and engulfment of country rocks are locally conspicuous.

Lopoliths are of very great economic importance. Because of the Cr, Pt, Ni, V, Cu, and Sn deposits, the Bushveld Complex has been described as the largest repository of magmatic ore deposits in the world. The Sudbury intrusion is the world’s principal source of Ni production: it also supplies a nearly equal amount of Cu, and quantities of precious metals (Au, Pt, Pd, Ir). The Duluth and Stillwater intrusions contain major resources of lower-grade Ni and Cu sulfides, precious metals (Au, Pt, Pd, Ir), and chromite. The Great Dyke of Zimbabwe has the western world’s largest deposits of metallurgical-grade chromite.

Accumulations of Fe, Ni, and Cu sulfides, with associated precious metals, typically form irregular, inclusion-rich, mafic layers at the lower contact of the intrusions, e.g., Sudbury, Duluth, Stillwater, Insizwa. Concentrations of chromite and other ore minerals form extensive regular layers within the complexes, e.g., chromite and Pt of the Bushveld, Great Dyke, and Stillwater bodies. Sulfide Ni is also a significant by-product of the Bushveld Pt mining.

The primary layering of lopoliths may be transected by a variety of later, minor intrusions, ranging from quartz veins and aplitic

dykes to dunitic pipes; the latter, in the Bushveld Complex, may carry rich but erratic Pt mineralization.

In recent years continuing studies of the major lopoliths have shown that they do not conform to Grout's or even Wilson's simplified models; this is reflected in a revised nomenclature: e.g., Duluth Complex, Bushveld Complex. Multiple differentiated intrusions, not simply multiple pulses of magma into a single lopolithic chamber, have been proposed for both the Duluth (see Phinney et al., in Sims and Morey, 1972) and Bushveld (Visser and von Gruenewaldt, 1970) intrusions, while Irvine et al. (1983) have highlighted the role of double-diffusive convective magma mixing in relation to Pt-Pd concentrations. The important sulfide concentrations at Sudbury have been identified as separate, marginal intrusions (Naldrett and Kullerud, 1967; Souch et al., 1969; Naldrett et al., 1970).

Whereas the forcible emplacement of a sill may proceed by raising the superincumbent strata, and the relatively small Skaergaard intrusion could be accommodated by upward displacement of a conical segment of crustal rocks, it is not easy to visualize similar processes for the great lopoliths. The suggestion by Dietz (1964) that the Sudbury structure was initiated by meteorite impact has been strongly supported by subsequent studies (see Guy-Bray, 1972; Pye et al., 1984): certainly a large impact crater produces not only a cavity but also a great volume of subjacent broken and fractured rock of crudely conical form. Unless something has been removed, there is no obvious reason why the structures should be synclinal.

Lopoliths are characteristically formed in stable cratonic or platform regions of the crust, and good reasons for the occurrence of huge volumes of magma in such environments are not often presented. The magmas are typically tholeiitic, having none of the alkaline tendencies of intrusions associated with deep rifting. On the other hand, they are localized, unlike the widespread plateau basalt eruptions associated with shallower crustal tension. It can be argued that major meteorite impacts could provide both a localizing mechanism and a trigger for generation of magma from the uppermost mantle. The Precambrian age of the largest impact scars also agrees with the age of the majority of lopoliths. However, except at Sudbury the connection is merely speculative.

J. V. GUY-BRAY

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Cross-references: *Astrobleme*; *Bushveld Complex*; *Emplacement—modes*; *Igneous intrusions—structural behavior*; *Intrusion—layered*; *Sill*.

LUNAR MINERALOGY—See Vol. IVB

LUNAR PETROLOGY

Among the terrestrial planets the Moon represents the least complex example of an evolved body. The very simplicity of the Moon's history makes it possible to pose a problem, perform an experiment on a limited number of samples and have some confidence in the generality of the observations. Critical to the petrological, geochemical, and isotopic studies of the Moon are the virtual lack of H₂O and CO₂, and of evidence for tectonic processes other than those driven by impact. Chemical fractionation on the Moon is assumed to take place almost exclusively by igneous processes (crystal settling and partial melting), and these processes never produced continental-type rocks



FIGURE 1. Lick Observatory photograph L4 of the full Moon; Albedo differences between maria, highlands, and fresh ejecta are well displayed.

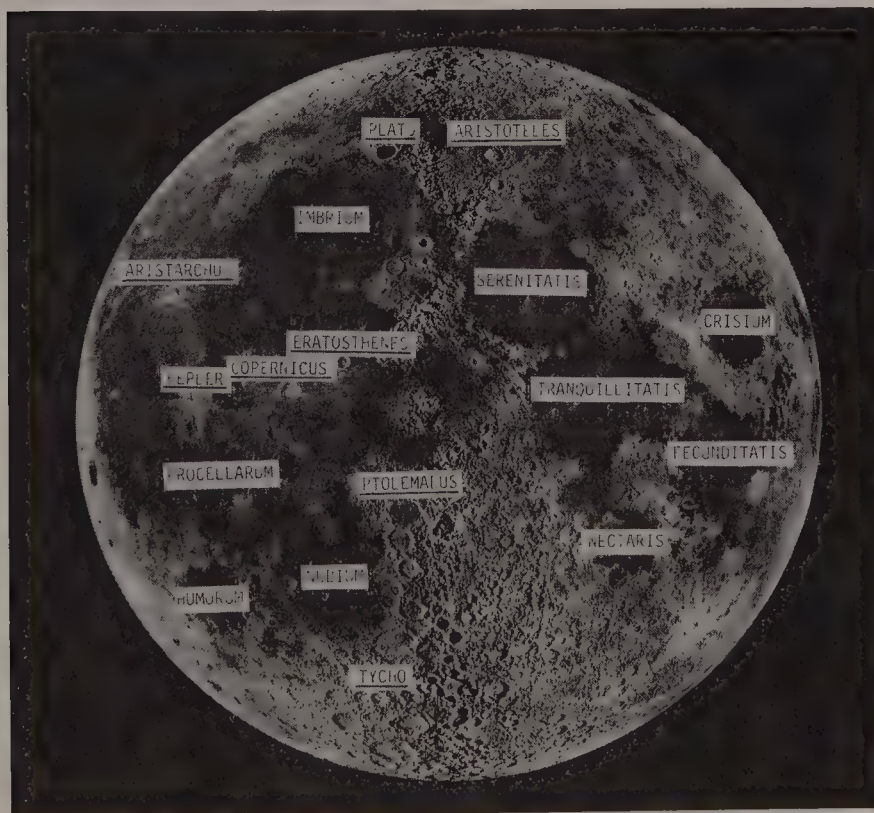


FIGURE 2. Lick Observatory photograph L9—a composite of two half-Moon photographs. The high density of craters in the highlands shows up well in a pole-to-pole band in the center of the Moon's nearside. Underlined labels refer to craters directly beneath the label; other labels refer to mare basins.

TABLE 1. Bulk lunar properties

Mean density	$3.344 \pm 0.004 \text{ gm/cm}^3$
Mean radius	1,738.09 km
C/MR^2	0.392 ± 0.003
Mean Earth–Moon distance	384,402 km
Surface gravity	162 cm/sec^2
Central pressure	42 kb
Seismic energy release	$<10^{15} \text{ ergs/hr}$
Surface heat flux	$2 \mu\text{W/cm}^2$
Magnetic dipole moment	$<4.4 \times 10^{13} \text{ G/cm}^3$

such as granites or andesites. The lack of H_2O and an atmosphere precludes the kinds of chemical migration thought to occur in metamorphic terranes on Earth. Although impact-induced fractionation by volatilization has been suggested, it has not been unequivocally demonstrated to operate in any environment other than the upper few meters of lunar soil, and is not relevant to the kilometer-scale and larger-scale impacts that dominate the Moon's cratering record.

General properties of the Moon have been known for many years. The results of the Apollo programme, however, have enormously amplified our knowledge of the Moon as a planet (Table 1).

Lunar regions

Lunar terrain is divided into two major regions (Fig. 1): dark maria and light highlands. Maria are topographically low, and fill circular and irregular basins or depressions. Mafic basalts (marebases) of several different chemical types cover most of these plains. The highlands stand about a kilometer above the maria and are heavily marked with circular features; these are mainly meteorite impact craters, which range in size from meters across, to large circular basins hundreds of kilometers across (Fig. 2). These areas consist of a mixture of mostly feldspar-rich (anorthositic) plutonic material and minor amounts of an Al- and trace element-rich basaltic material known as KREEP.

Knowledge of the Moon's interior comes from limited data modeled with the Moon's bulk density and moment of inertia. There is some evidence from a single large seismic signal (impact of the third stage of a Saturn IVB booster) that suggests a high-density, low-velocity core, forming $<5\%$ of the Moon's mass. The Moon has a crust with a velocity structure consistent with anorthositic composition, about 60 km thick on the Earth-facing side, and some 100 km thick on the opposite side. The mantle occupies most of the Moon's volume and mass.

Bulk composition

Based on the study of a few samples from the Moon's outer layer (Apollo and Luna program samples), some broad generalizations may be made about bulk compositions. Relative to Earth, the Moon is depleted in volatile and siderophile elements. The total lack of any trace of water or hydrated minerals in the lunar samples attests to volatile depletion. The Moon has systematically lower K/U ratios than any other planetary object (Fig. 3), demonstrating relative depletion in K (a volatile element) over U (a more refractory element). Figure 4 shows the ratios of various elements in lunar and terrestrial basalts plotted against their condensation temperature. Since basalts derive from partial melting of mantle rocks, this plot suggests that the Moon's mantle is depleted in volatile elements compared to the Earth's.

The low bulk density of the Moon and its limited core size (if one exists at all) suggest that the Moon is depleted in Fe and the other elements present in the Earth's core. Analyses of basalts indicate the Moon's mantle to be depleted in trace siderophile elements (Au, Ni, Pt, Pd, Ir, Re, and Os).

Mare basalts

The basins of the Moon are filled with Fe-rich basalts that display a wide range of textures from vitrophyric, porphyritic, and subophitic to intersertal. Vitrophyric basalts have olivine and/or pyroxene phenocrysts in a glass or devitrified glass matrix. Porphyritic basalts have similar phenocrysts set in a holocrystalline matrix that ranges from large, single crystals of poikilitic plagioclase to subophitic or feathery intergrowths of plagioclase and pyroxene. Subophitic basalts contain tabular to equant plagioclase intergrown with subhedral pyroxene and olivine crystals. Intersertal basalts contain anhedral crystals of plagioclase, pyroxene, olivine and ilmenite set in a glassy mesostasis. All are fine-grained by terrestrial standards, the largest grain-size being $<1 \text{ mm}$, except for phenocrysts of olivine and pyroxene, which may exceed several centimeters. No plagioclase phenocrysts have been observed.

Mare basalts contain calcic plagioclase, clinopyroxene, olivine, ilmenite, and a suite of accessory minerals. Plagioclase contents range from 15 to 35%; compositions range from An_{95} to An_{75} . Plagioclase crystals are not strongly zoned, but they consistently display Na-enrichment on their rims. Pyroxene is the most abundant mineral, being pigeonite and/or subcalcic augite; these are relatively Fe-rich, and display extreme Fe-enrichment. Many large pyroxene crystals, especially the phenocrysts,

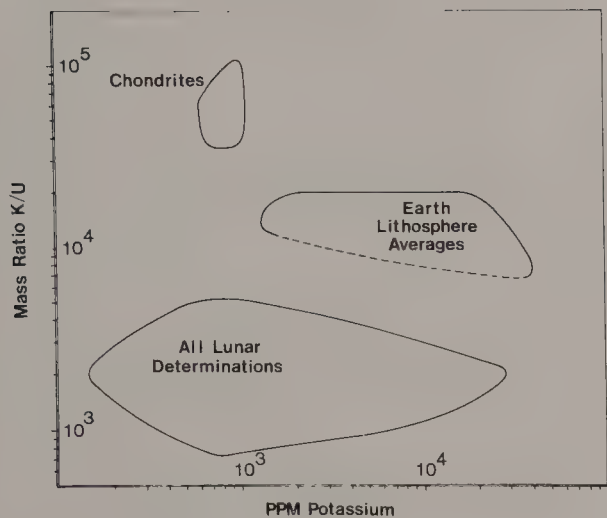


FIGURE 3. K/U mass ratio vs. p.p.m. K.

are complexly zoned. Two types of zoning have been described: sector zoning in which crystal growth in different crystallographic directions has resulted in different compositions; and growth zoning, in which, for example, a core of pigeonite is overgrown by a rim of sub-calcic augite, which itself is overgrown by a second rim of Fe-rich pigeonite (Fig. 5). Fe-enrichment may be so extreme that pyroxferroite (an Fe-rich pyroxenoid—Fig. 5) formed, as a rim on pyroxenes. Olivine compositions range from Fo₈₀ to Fo₃₀; in addition, fayalite occurs as a late-stage phase in some basalts. Some basalts that show evidence of olivine accumulation may

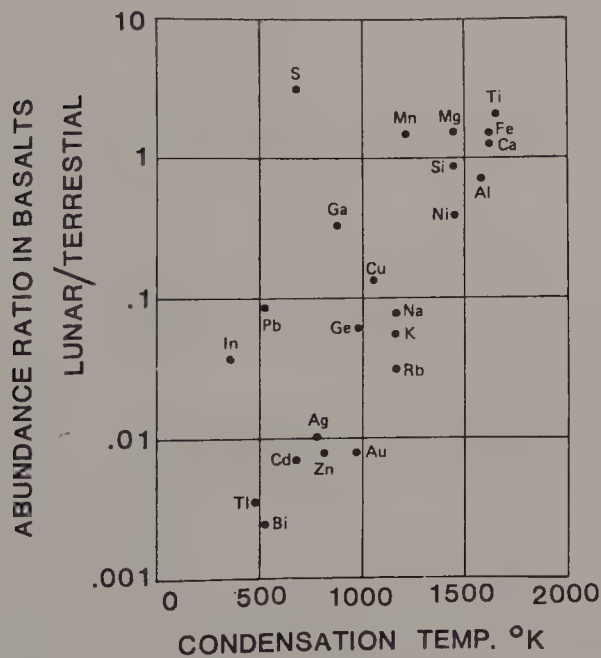


FIGURE 4. Relative abundance of elements in lunar and terrestrial basalts as a function of each element's volatility as measured by the temperature at which each element will condense from a nebula of solar composition. (After Wood, 1979.)

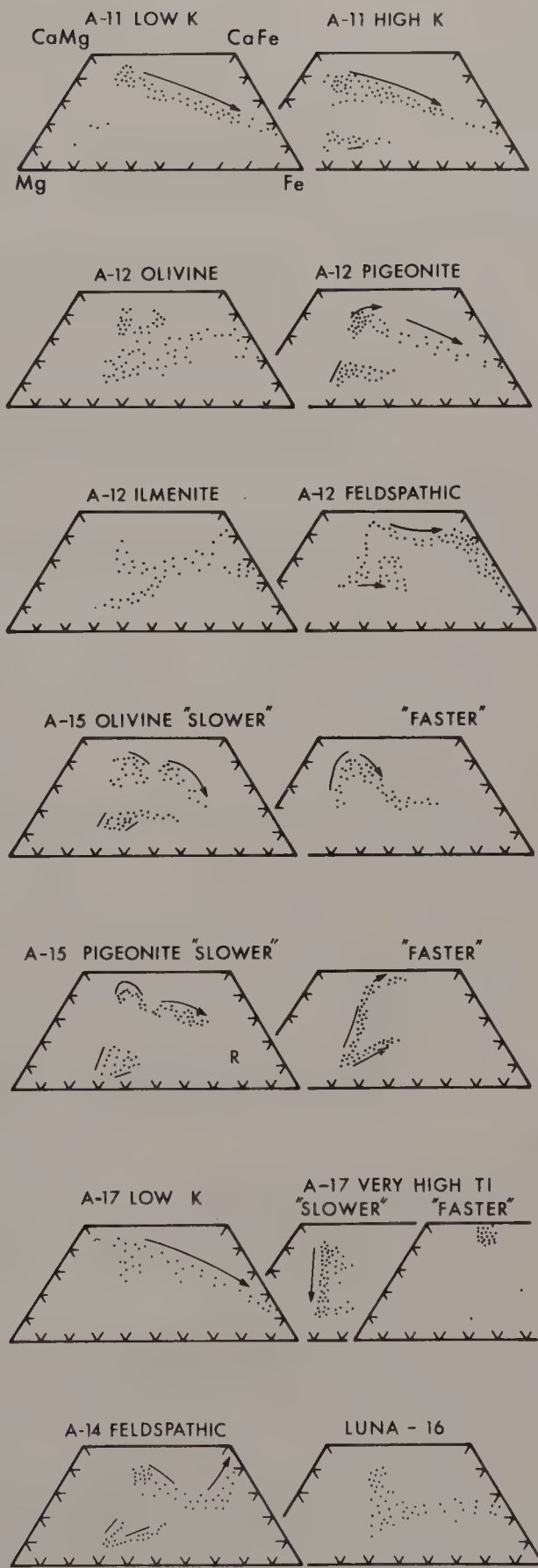


FIGURE 5. Schematic pyroxene composition trends shown on the pyroxene quadrilateral for various mare basalt groups. (After Papike et al., 1976.)

TABLE 2. Chemical analyses of mare basalts

Sample	Low Ti series					High Ti series				
	Apollo 12 oliv-pig	Apollo 12 ilmenite	Apollo 12 feldspathic	Apollo 15 quartz	Apollo 15 olivine	Apollo 15 green glass	Apollo 11 high K	Apollo 11 low K	Apollo 12	Apollo 17 orange glass
SiO ₂ ^a	46.42	44.44	46.56	47.94	44.90	45.2	40.37	40.37	39.48	38.8
TiO ₂	2.14	4.59	3.31	1.87	2.41	1.1	11.77	10.18	11.77	8.8
Al ₂ O ₃	9.15	9.81	12.53	9.49	8.81	15.1	8.84	10.40	9.38	6.4
Cr ₂ O ₃	0.52	0.42	0.27	0.47	0.57	0.4	0.36	0.29	0.39	0.7
FeO	21.33	20.77	17.99	20.76	22.41	13.7	19.28	18.73	18.84	22.2
MnO	0.29	0.27	0.27	0.29	0.30	0.2	0.23	0.27	0.28	0.3
MgO	9.48	8.55	6.71	8.55	10.41	12.14	7.56	6.92	7.97	17.4
CaO	10.25	10.80	11.62	10.63	9.87	11.11	10.59	10.63	11.11	7.4
Na ₂ O	0.30	0.31	0.66	0.32	0.28	0.4	0.52	0.40	0.41	0.4
K ₂ O	0.05	0.07	0.07	0.06	0.05	0.1	0.31	0.07	0.06	0.08
P ₂ O ₅	0.08	0.12	0.14	0.08	0.08	0.09	0.17	0.09	0.05	0.04
S	0.07	0.09	0.06	0.07	0.07	0.06	0.22	0.16	0.17	0.07
Li ^b	—	—	—	—	—	—	—	—	9.9	10.7
Rb	0.8	0.9	0.48	0.9	0.7	0.34	5.5	0.74	0.65	1.11
Sr	102	143	185.8	108	95	—	166	174	193	209
Ba	54	74	120	62	48	17	308	99	83.3	76.4
Sc	—	—	—	—	—	—	83	90	80.1	—
La	5.4	6.3	11.80	5.8	4.9	1.4	27	10	6.74	6.25
Ce	14	18	29.1	15	14	3.8	76	34	24.0	19.0
Nd	10	15	22.0	11	10	2.2	63	31	26.0	17.8
Sm	3.4	5.9	7.57	3.6	3.2	0.76	21	12	10.9	6.53
Eu	0.93	1.26	1.97	0.97	0.89	0.21	2.2	1.9	2.16	1.80
Gd	4.7	8.3	10.10	4.9	4.6	0.91	27	17	16.7	8.52
Tb	—	—	1.61	—	—	0.15	—	—	—	—
Dy	5.2	9.9	9.73	5.5	4.8	1.1	33	19	19.2	9.40
Er	3.0	5.5	5.0	3.3	2.7	0.8	20	12	11.4	5.10
Yb	2.3	5.3	4.3	2.7	2.2	0.93	18	11	10.4	4.43

Lu	0.32	0.74	0.69	0.33	0.32	0.14	2.6	1.6	1.43	0.61
Zr ^b	89	121	—	95	83	22	—	—	—	—
Hf	—	—	—	—	—	0.42	18	12	9.6	—
Th	—	—	—	—	—	0.08	—	—	—	—
U	—	—	—	—	—	0.02	—	—	—	—
Ir ^c	0.10	0.07	0.04	0.006	0.02	0.22	0.02	0.006	—	0.214
Re	—	—	—	0.004	0.002	0.020	—	—	0.019	0.0553
Au	0.03	0.08	—	0.03	0.026	0.188	0.03	0.04	0.029	1.07
Co ^b	—	60	25	42	51	72	28	16	17.9	—
Ni ^b	22	7	—	10	45	170	—	—	1.5	70
Sb ^c	—	—	—	0.02	0.13	0.12	—	—	—	25.3
Ge	—	—	—	5.7	9.4	37	—	—	3.5	191
Se	170	170	120	140	150	69	—	—	—	460
Te	30	30	—	2.7	3.0	3.3	8	13	—	49
Ag	0.6	1.0	—	0.90	1	8.9	1	2	—	75
Bi	0.3	0.4	—	0.16	1.3	0.38	—	—	—	1.53
Zn ^b	0.8	0.7	—	0.14	1.5	19	1.7	1.3	1.7	200
Cd ^c	—	1.2	—	1.6	1.7	46	6	5	1.9	260
Tl ^c	0.3	0.35	—	0.40	0.32	1.13	0.9	0.3	—	9.9
Norm (wt.%)										
Quartz	—	—	0.80	1.01	—	—	1.93	2.67	1.07	—
Orthoclase	0.30	0.41	0.41	0.35	0.30	0.59	1.83	0.41	0.35	0.47
Albite	2.54	2.62	5.58	2.71	2.37	3.38	4.40	3.38	3.47	3.38
Anorthite	23.47	25.17	31.02	24.28	22.64	39.11	20.87	26.36	23.58	15.43
Diopside	22.58	23.27	21.72	23.53	21.66	12.70	25.24	21.40	25.64	17.13
Hyperssthene	40.71	35.64	33.60	43.69	33.80	23.18	22.47	24.46	22.60	13.79
Olivine	5.40	3.50	—	—	13.74	17.70	—	—	—	34.48
Ilmenite	4.06	8.72	6.29	3.55	4.58	2.09	22.35	19.33	22.35	16.71
Apatite	0.17	0.26	0.31	0.17	0.17	0.20	0.37	0.20	0.11	0.09
Chromite	0.78	0.62	0.40	0.69	0.84	0.59	0.53	0.43	0.57	1.03

^a SiO₂-S in %.

^b Li-U, and Co, Ni, Zr in p.p.m.

^c Ir-Au, Sb-Bi, and Cd, Tl in p.p.b.

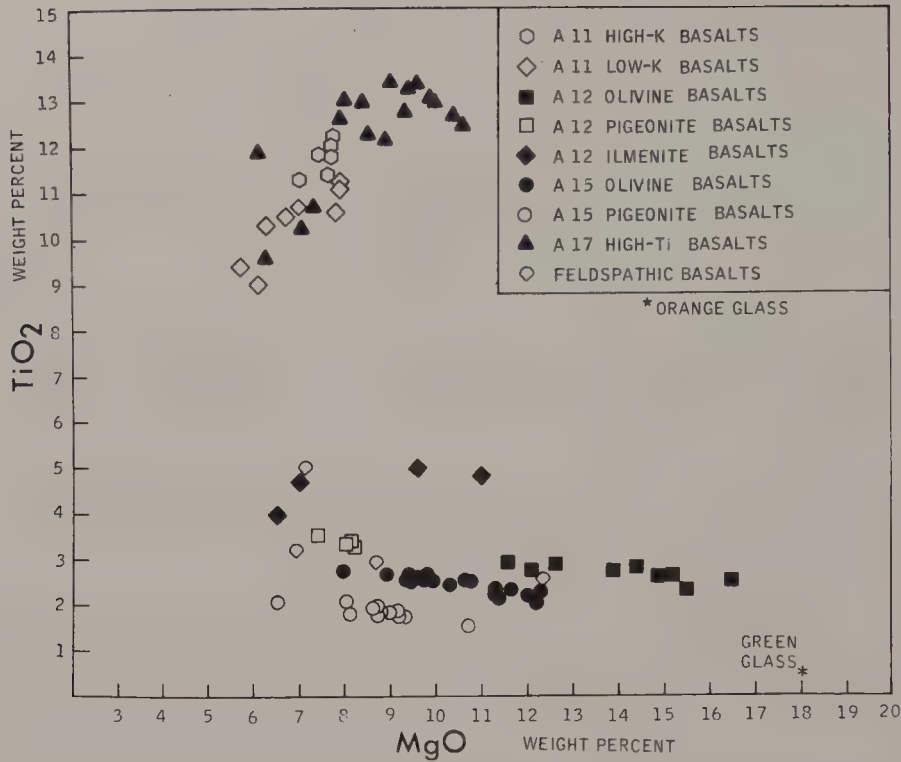


FIGURE 6. TiO_2 vs. MgO plot for mare basalts; green and orange glasses are pyroclastic deposits found at Apollo 15 and Apollo 17 sites. (After Papike et al., 1976.)

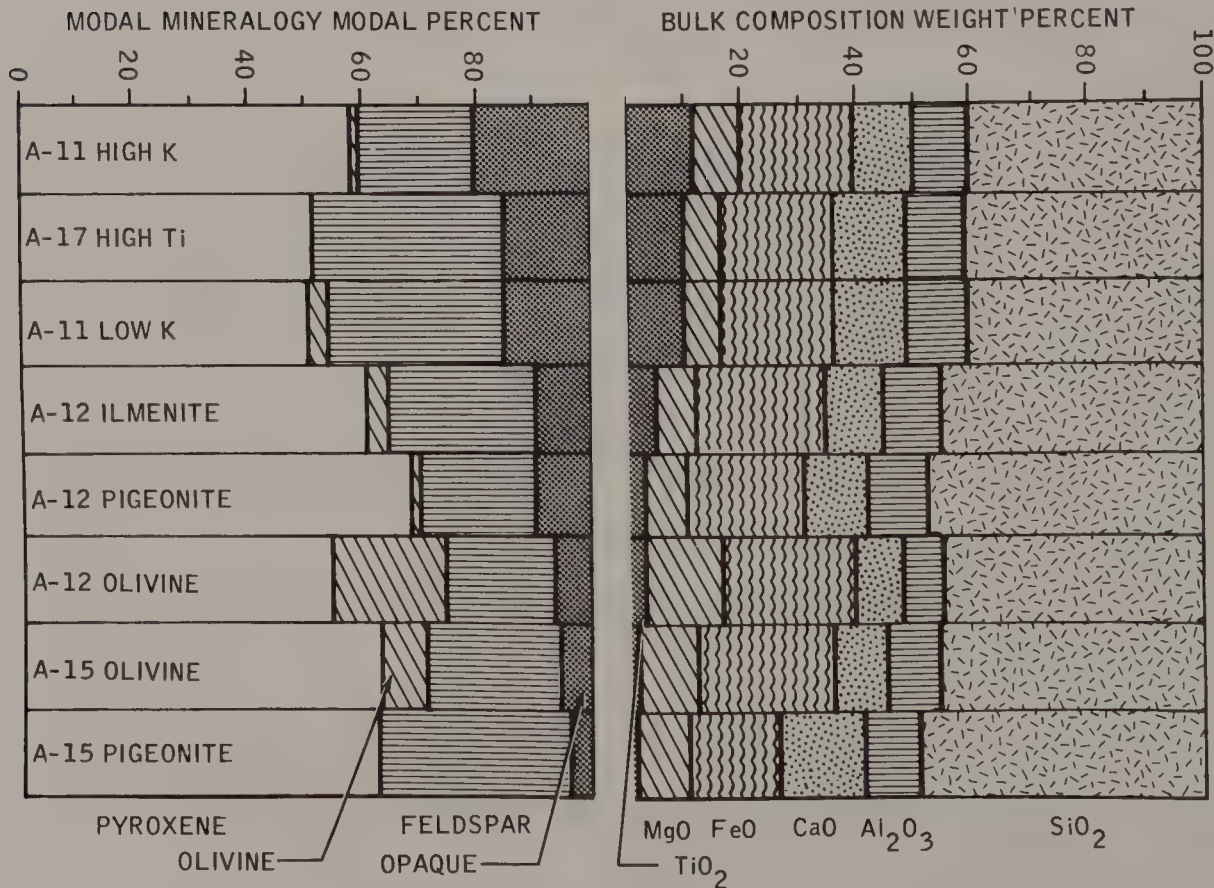


FIGURE 7. Schematic representation of the modal mineralogy and bulk chemical composition for the various mare basalt groups. (After Papike et al., 1976.)

have an olivine content as high as 30%. Ilmenite is the major opaque mineral making up 5–15% of mare basalt. Many basalts contain glassy mesostasis and/or a silica phase. The mesostasis commonly consists of high-Si material that formed by liquid immiscibility from the fractionating basalt. The silica phase is most commonly cristobalite, or more rarely tridymite. Accessory minerals include spinels in the series chromite to ulvöspinel, Fe–Ni metal, triolite (FeS), apatite and/or whitlockite, armalcolite, and other Ti and Zr silicates and oxides. New minerals found in mare basalts include pyroxferroite and armalcolite (a TiO_2 -rich oxide).

Oxygen fugacities are very low, generally being below the iron–wustite buffer by about one-half a log unit (values of about 10^{-12} at liquidus temperatures of $1,200^\circ\text{C}$).

Chemically, the mare basalts are rich in Ti and Fe, poor in Al, very poor in alkalis, and have a high Ca/Al ratio. They are depleted in siderophile trace elements and enriched in

lithophile trace elements (Table 2; Figs. 6, 7, 8). The lithophile-element abundance pattern (relative to chondrite—Fig. 8) shows that these elements are enriched by factors of 20–100. The exceptions are Eu and Sr, elements that may occupy lower valence states. The relative depletion in these elements is attributed to equilibrium between plagioclase and the basaltic liquid, where Eu and Sr are fractionated into plagioclase.

Mare basalts have been classified into one or more well-defined groups at each landing site. Individual samples within each group display a range of major and trace constituents that are explained by near-surface fractionation (low pressure crystal–liquid equilibrium) involving olivine, ilmenite or armalcolite, Cr-spinel, and/or low-Ca pyroxene.

Ti content. High-Ti basalts contain 10–13 wt.% TiO_2 ; low-Ti basalts contain 1–5 wt.% TiO_2 .

Alkali (K) content. Most mare basalts have

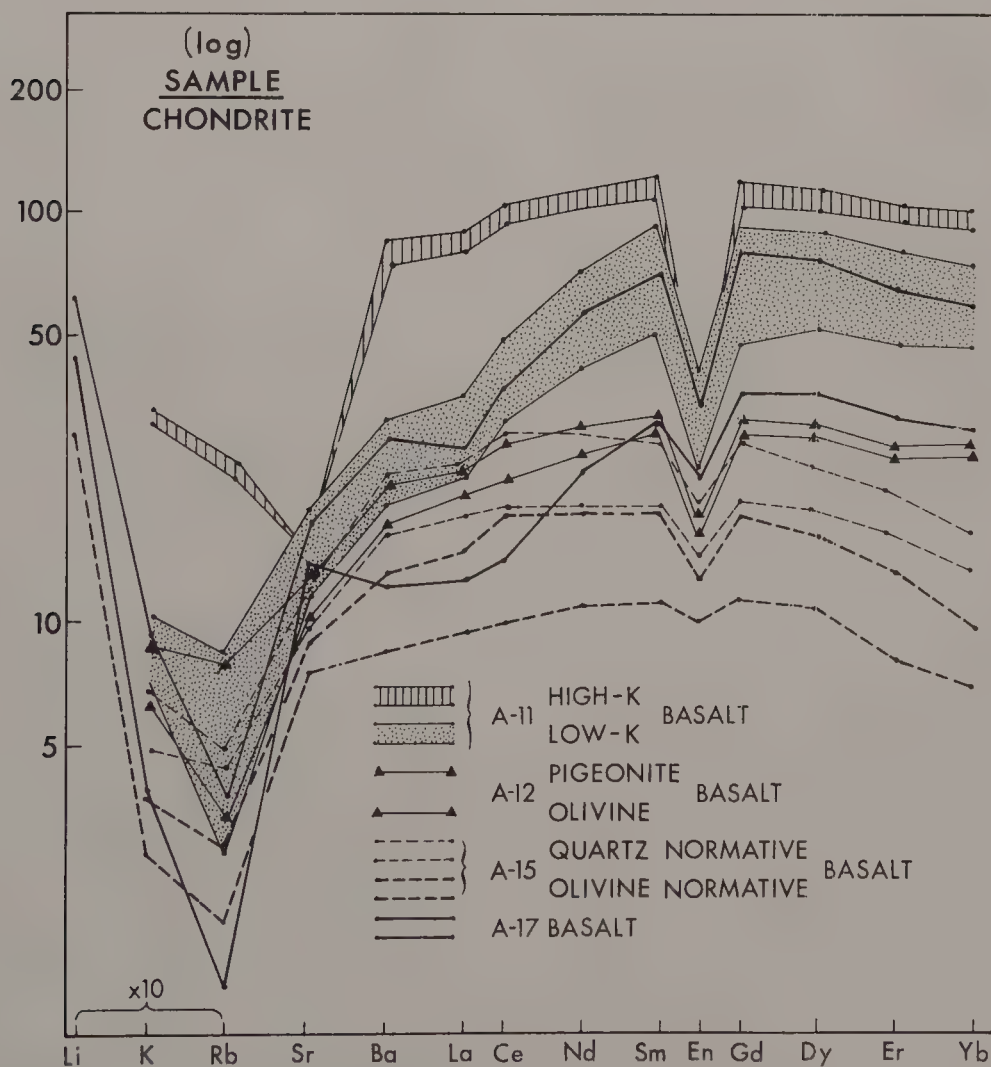


FIGURE 8. Lithophile trace-element concentrations in mare basalt chemical groups plotted with respect to the concentrations observed in chondritic meteorites.

low alkali contents. Using K as a monitor, low-K basalts contain <0.1 wt.% K_2O . High-K basalts contain 0.3 wt.% K_2O , occur only at the Apollo 11 site, and display a distinctive intersertal texture.

Al content. Most mare basalts contain <11 wt.% Al_2O_3 and >18 wt.% FeO. Several small groups of high-Al or feldspathic basalts have higher Al and lower Fe contents; they occur only in the Apollo 12 and Luna 16 sample collections.

Fractionation trends. Several groups of basalt have been identified because they define separate trends on variation diagrams (e.g., TiO_2 vs. MgO —Fig. 6). The Apollo 12 olivine–pigeonite and ilmenite trends, and the Apollo 15 quartz and olivine trends, are examples.

There are several significant and systematic differences between terrestrial and lunar basalts. The most striking, which can be observed in thin section, is that lunar basalts lack water. There are no amphiboles, or micas, no sericitization of feldspars, no chloritization of mafic minerals, and no leucoxene rims on ilmenite. The second most striking difference between terrestrial and lunar basalts is that the latter are depleted in volatiles, including major elements K and Na, and trace elements such as Pb and Bi. Siderophile trace elements such as Ni and Au are also deficient in lunar basalts. Whereas volatiles and siderophile elements in the Moon are depleted relative to the Earth, lithophile elements, such as REE, are enriched.

Crystallization ages obtained from Rb–Sr and Nd–Sm internal isochrons and from $^{39}Ar/^{40}Ar$ gas release plateaus lie between 3.15 and 3.9 Ga. Photogeological studies of relative ages of various parts of mare surfaces (using crater counts and morphologies) suggest that the Moon has mare basalt flows that are significantly younger than any sampled flow. These may be as young as 2.0 Ga. There is no evidence for more recent volcanic activity.

The origin of mare basalt magma is not well established, as no petrogenetic scheme is able to satisfy all geochemical and geophysical constraints. Experimental petrology suggests that the low-Ti suite formed by partial melting of an olivine pyroxenite mantle at depths of 200–500 km. Experimental petrology suggests that the high-Ti suite formed by partial melting of an intrinsically Ti-rich part of an olivine pyroxenite mantle at a depth of several hundred kilometers. Because these source regions do not contain plagioclase, the depletion in Eu and Sr must have already been present when the partial melting occurred. Sr, Nd, and Pb isotope systematics indicate that these source regions had a prior differentiation history that ended at

about 4.4 Ga. The depletion in Eu and Sr must have occurred during this early stage of differentiation. Alternatively, the high-Ti suite magmas may have originated by an undefined mixing or hybridization process in which low-Ti magma combined with the Ti-rich residue of an earlier differentiation.

Rocks of the lunar highlands

Igneous rocks and polymict breccias occur in the lunar highlands. The igneous rocks are of two types: (1) plutonic cumulates that are relics of the original crust, and (2) basaltic rocks enriched in K, REE and P (KREEP), derived from the liquid residue from the fractionation that produced the lunar crust. The breccias are the product of meteorite impacts.

A major issue in highland petrology is to distinguish the igneous rocks from the polymict breccias. This problem is important because petrogenetic arguments are critically dependent on whether a particular rock is igneous and therefore part of the Moon's fractionation history, or whether that rock is a breccia and therefore a mixture of unknown, preexisting materials.

Breccias are identified petrographically because they display a clast–matrix structure with the clasts defining a polymict assemblage, and chemically because they display enriched abundances of siderophile trace-elements in which the moon as a whole is depleted. As an example, the mare basalts contain 0.01–0.1 p.p.b. Au, whereas the overlying regolith contains 1–2 p.p.b. This enrichment is due to contamination by about 1 or 2% meteorite material. Breccias were formed by meteorite impact, so they also contain a few percent meteorite material, and mimic the regolith with enriched abundances of siderophile trace elements. Polymict breccias typically contain 0.3–20 p.p.b. Au. The chemical criteria for the identification of breccias is important because many igneous rocks from the highlands have been crushed and superficially resemble breccias. Such crushed igneous rocks are not polymict, because they are chemically pristine.

Chemical compositions and CIPW norms for highland igneous rocks and polymict breccias are given in Tables 3 and 4, and plotted on the silica (quartz)–olivine–plagioclase pseudoternary diagram (Fig. 9). This system faithfully represents the lunar highlands, because these rocks have small contents of Ca-pyroxene, alkali feldspar, and Fe–Ti oxides. Lithophile trace-element abundances with respect to chondrite compositions are illustrated in Fig. 10.

Plutonic cumulates are coarse-grained relative to other lunar materials (>1 mm). Some of these rocks display gabbroic textures with

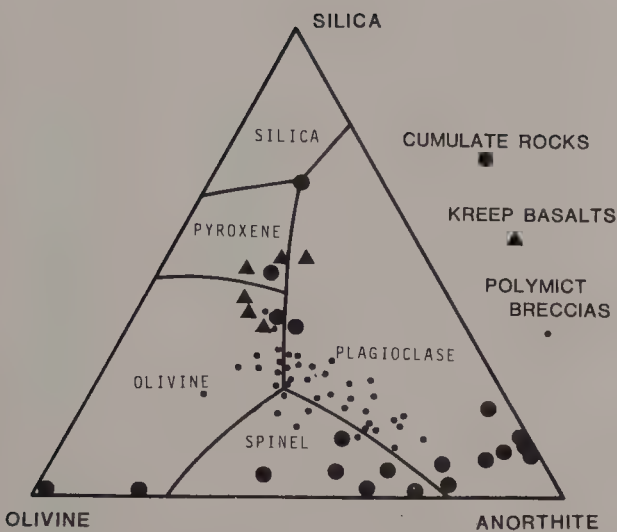


FIGURE 9. Major-element compositions for rocks from the lunar highlands; data are plotted on the liquidus diagram for silica(quartz)–plagioclase–olivine determined for general lunar compositions with $Mg/(Mg + Fe) = 0.3$ mol. %.

oscillatory zoned plagioclase and complexly exsolved pyroxenes. Most display granulitic textures with smooth crystal boundaries meeting at 120° triple points, and with homogeneous mineral compositions with all grains of each phase having the same composition. Most have

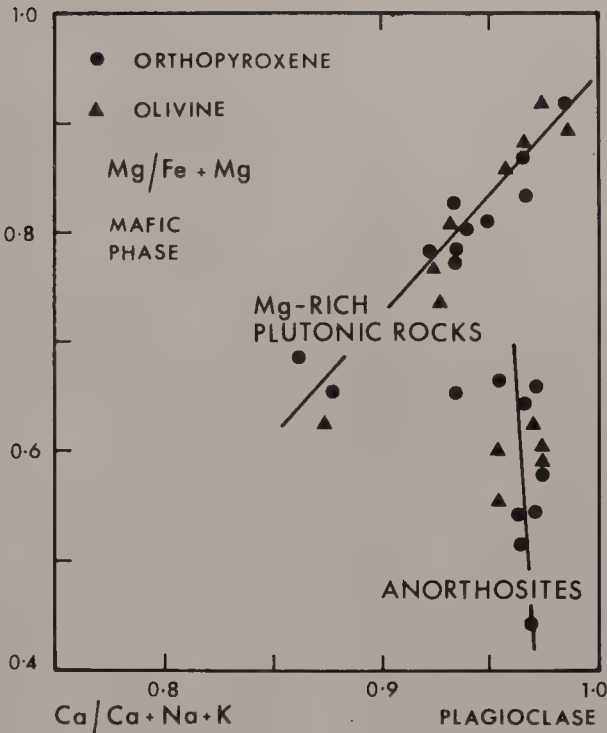


FIGURE 11. Plot of mineral chemistry of coexisting minerals in plutonic rocks; one trend (Mg-rich plutonic rocks) is similar to trends observed for magmatic differentiation on Earth; the other trend (anorthosites) is unique to the Moon. (After Warner and Bickel, 1978.)

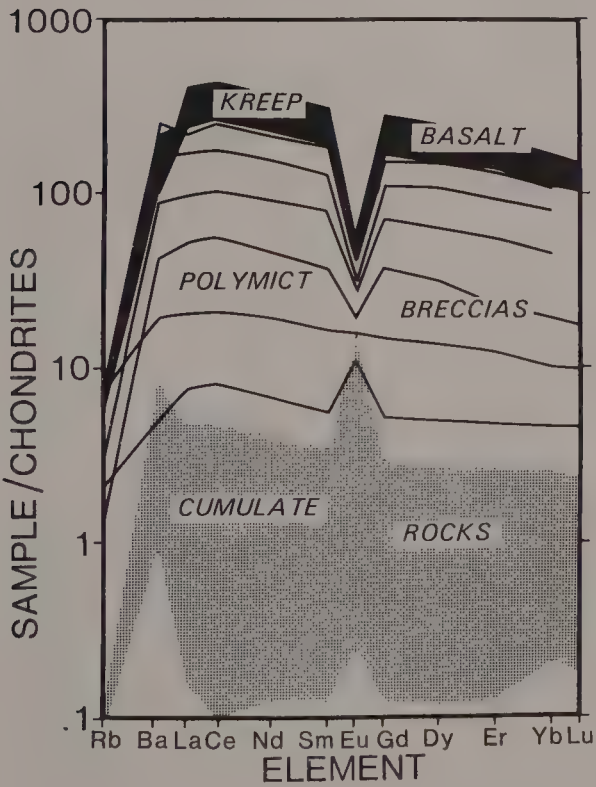


FIGURE 10. Lithophile trace-element concentrations in rocks from the lunar highlands plotted with respect to the concentrations observed in chondritic meteorites.

been crushed during excavation by impact and are now cataclastic rocks.

Mineralogically, these cumulates are simple rocks, being a mixture of calcic plagioclase, magnesian olivine and magnesian orthopyroxene, with minor magnesian augite, Mg–Al spinel and traces of Fe–Ti–Cr oxides, Fe-metal, troilite, and phosphates (Fig. 11). There is one trend of increasing Fe in the mafic minerals and Na in the feldspar that is typical of magmatic fractionation observed in terrestrial layered intrusions. A second trend, with Fe-enrichment in mafic minerals and constant plagioclase composition, is defined by anorthosite samples. This simple mineralogy, combined with low and unfractionated lithophile trace-element patterns (Fig. 10) and Mg-rich major-element compositions that plot far from any cotectic thermal valley (Fig. 9), suggest that these plutonic rocks are a series of cumulates. They are anorthosite, troctolite, dunite, spinel troctolite, gabbro, and norite.

A troctolite and a dunite sample contain diopside–enstatite–spinel symplectite along olivine–plagioclase boundaries; the symplectite formed by reaction of olivine and plagioclase at a depth of 10–30 km and has been dated at 4.3–4.6 Ga by Rb–Sr and Nd–Sm internal

TABLE 3. Chemical analyses of pristine rocks from the lunar highlands

Sample	Anorthosite (60025)	Dunite (72415)	Feldpathic troctolite (76535)	KREEP (15386)	KREEP (15382)
SiO ₂ ^a	44.3	39.9	42.9	50.83	—
TiO ₂	0.02	0.03	0.05	2.23	2.17
Al ₂ O ₃	35.2	1.5	20.7	14.77	14.9
Cr ₂ O ₃	0.3	0.3	0.1	0.35	—
FeO	0.5	11.3	5.0	10.55	9.2
MnO	0.02	0.1	0.07	0.16	—
MgO	0.2	43.6	19.1	8.17	7.4
CaO	19.2	1.1	11.4	9.71	7.1
Na ₂ O	0.5	<0.02	0.2	0.73	0.85
K ₂ O	0.03	0.0	0.03	0.67	0.73
P ₂ O ₅	0.003	0.04	0.03	0.70	—
S	—	0.01	—	—	—
Li ^b	—	4.9	3.0	27.2	—
Rb	<0.1	0.066	0.24	18.5	16.1
Sr	—	2.24	114	187	195
Ba	—	3.27	32.7	837	793
Sc	—	—	—	—	—
La	0.28	0.05	1.51	83.5	79.5
Ce	0.65	0.07	3.81	211	212
Nd	0.42	0.07	2.30	131	127
Sm	0.072	0.022	0.61	37.52	35.2
Eu	1.04	0.016	0.73	2.72	2.77
Gd	—	0.030	0.73	45.4	42.9
Tb	—	—	—	—	—
Dy	0.19	0.035	0.80	46.3	45.7
Er	0.05	0.04	0.53	27.3	28.1
Yb	0.048	0.045	0.56	24.4	24.0
Lu	0.006	0.008	0.079	—	3.43
Zr ^b	—	3.0	24	970	—
Hf	0.02	0.015	0.52	—	—
Th	—	—	0.16	10	—
U	—	<0.005	0.056	0.061	3.72
Ir ^c	0.0057	0.0052	0.0054	0.22	0.0132
Re	0.0016	0.0048	0.0012	—	0.0089
Au	0.0074	0.255	0.0025	—	0.0033
Co ^b	—	—	—	23	—
Ni ^b	0.3	149	44	12.5	18
Sb ^c	0.035	0.47	0.014	—	0.17
Ge	2.30	29.8	1.70	61	47.2
Se	21.7	4.9	4.1	—	72
Te	65	<0.36	0.28	—	1.0
Ag	0.22	0.25	0.12	—	0.44
Bi	3.58	0.41	0.037	—	0.29
Zn	0.17	2.1	1.2	3.5	2.6
Cd ^c	7.25	0.37	0.60	10	86.6
Tl ^c	26	0.049	0.12	—	3.2
Norm (wt.%)					
Quartz	0.07	—	—	7.62	—
Orthoclase	0.18	—	0.18	3.96	—
Albite	4.11	—	1.69	6.18	—
Anorthite	93.72	4.09	55.50	35.50	—
Diopside	1.30	0.90	0.70	7.39	—
Hypersthene	—	2.02	3.64	32.40	—
Olivine	0.55	90.28	37.56	—	—
Ilmenite	0.04	0.06	0.09	4.24	—
Apatite	—	0.09	0.07	1.53	—
Chromite	0.04	0.44	0.15	0.52	—

^aSiO₂-S in %.
^bLi-U, and Co, Ni, Zr in p.p.m.
^cIr-Au, Sb-Bi, and Cd, Tl in p.p.b.

TABLE 4. Chemical analyses of polymict breccias

Sample	12013 light	12013 dark	14301	14310	61016	62235	68415	76315	79215
SiO ₂ ^a	—	—	48.26	47.2	43.82	47.05	45.4	46.21	44.8
TiO ₂	0.3	3.3	2.06	1.24	0.69	1.19	0.32	1.50	0.3
Al ₂ O ₃	10.1	14.6	16.52	20.1	25.06	18.88	28.63	18.14	27.6
Cr ₂ O ₃	0.25	0.162	0.21	0.18	0.11	0.19	—	0.20	0.098
FeO	14.0	13.70	10.29	8.38	4.97	9.45	4.25	8.95	4.15
MnO	0.15	0.166	0.14	0.11	0.05	0.13	0.06	0.12	0.054
MgO	8.50	9.0	9.98	7.87	10.48	10.0	4.38	12.02	5.84
CaO	4.60	9.7	10.29	12.3	14.31	11.8	16.39	11.32	16.3
Na ₂ O	1.12	1.45	0.84	0.63	0.36	0.42	0.41	0.60	0.549
K ₂ O	2.60	0.40	0.75	0.49	0.07	0.33	0.06	0.26	0.107
P ₂ O ₅	—	—	0.64	0.34	0.12	0.39	0.07	0.29	—
S	—	—	—	0.02	—	0.1	0.04	0.97	—
Li ^b	—	—	—	—	7.3	—	5.1	13.9	—
Rb	66.5	13.5	21.7	12.8	2.84	8.39	1.704	5.78	0.489
Sr	—	—	185	188	130	152	182.4	174	—
Ba	3760	390	959	617	160	530	76.2	337	—
Sc	25	28	—	—	6.6	16.5	—	—	7.14
La	56.5	135.7	71.8	56.4	16.7	60.3	6.81	31.6	2.5
Ce	151	347	201	144	46.0	160	18.3	82.3	6.5
Nd	74	215	121	87	—	94	10.9	52.7	—
Sm	19	59	34.7	24	6.9	25.2	3.09	14.8	1.03
Eu	2.14	3.89	2.69	2.15	1.38	1.93	1.11	1.95	0.77
Gd	—	—	40.3	28.1	9.5	31.7	3.78	18.8	—
Tb	5.2	12.5	—	—	1.5	5.56	—	—	0.23
Dy	33	75	46	32.7	9.7	35.0	4.18	19.1	—
Er	—	—	28	19.1	4.8	21.0	2.57	11.4	—
Yb	30	39	25	18.4	4.4	19.4	2.29	10.4	1.07
Lu	4.2	5.2	—	—	0.61	2.59	0.34	—	0.16
Zr ^b	690	2070	—	700	209	858	97.5	—	—
Hf	27	63	—	—	4.9	20.7	2.4	0.60	1.34
Th	47.5	24.5	—	11	1.6	7.83	1.26	5.69	0.21
U	14.2	10.4	—	—	0.46	2.08	0.32	2.52	0.190
Ir ^c	0.047	4.6	—	10.5	11.5	17.0	4.58	5.42	21.3
Re	—	—	—	1.02	1.28	1.7	0.434	0.507	1.9
Au	0.21	3.10	—	4.31	9.55	17.6	2.65	3.21	8.27
Co ^b	24	31	—	—	36.7	54.2	—	—	18.8
Ni ^b	—	—	—	64–410	515	830	165	256	255
Sb ^c	—	—	—	4.5	3.13	—	0.53	1.49	2.79
Ge	—	—	—	130	620	0.8	73	346	33
Se	54	99	—	120	181	260	98	100	176
Te	—	—	—	4	4.8	—	13.5	4.04	17.0
Ag	0.88	.4	—	10.8	21.4	—	4.8	0.84	1.10
Bi	0.46	.63	—	2.5	7.3	—	0.45	0.098	0.16
Zn	2.06	.10	—	2.3	0.84	11	4.8	3.1	2.3
Cd ^c	91	—	—	2.6	29.5	—	2.75	5.0	0.98
Tl ^c	18	—	—	10	45.3	—	0.49	0.31	0.41
Norm (wt.%)									
Quartz			0	0.16	—	—	—	—	—
Orthoclase			4.43	2.90	0.41	1.95	0.35	1.54	0.63
Albite			7.11	5.33	3.05	3.55	3.47	5.08	4.65
Anorthite			39.10	50.57	66.56	48.66	76.11	46.04	72.53
Diopside			6.68	6.91	2.98	6.18	3.92	6.80	6.74
Hypersthene			36.74	29.61	6.63	32.72	12.62	25.20	4.31
Olivine			—	—	18.68	3.38	2.75	11.19	10.22
Ilmenite			3.91	2.36	1.31	2.26	0.61	2.85	0.57
Apatite			1.40	0.74	0.26	0.85	0.15	0.63	—
Chromite			0.51	0.27	0.16	0.28	—	0.29	0.14

^a SiO₂–S in %
^b Li–U, and Co, Ni, Zr in p.p.m.
^c Ir–Au, Sb–Bi, and Cd, Tl in p.p.b.

isochrons. The ages, combined with the symplectite evidence for an origin at depth, suggests that the troctolite and the dunite, and perhaps other plutonic rocks, are samples of the original lunar crust. The occurrence of a clast of cordierite–spinel troctolite amongst regolith breccia suggests that spinel-bearing cumulates may be an important-constituent of the lower lunar crust (Marvin et al., 1989).

Many anorthosites display $^{39}\text{Ar}/^{40}\text{Ar}$ gas-release patterns that suggest crystallization (or reequilibration) ages of 3.9–4.3 Ga. However, $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are about 0.699, equal to the initial Sr isotope ratio. Ratios as unevolved as 0.699 suggest that the anorthosites separated from any significant amount of Rb within 0.1 Ga of the Moon's formation (i.e., ca. 4.5 Ga). The extreme age of these samples suggests that they formed part of the original lunar crust.

KREEP is a rare suite of highland igneous rocks with basaltic textures, characterized by pyroxene crystals with magnesian orthopyroxene cores and pigeonite and augite rims, intergrown with lath-shaped bent calcic plagioclase crystals with small amounts of Na-enrichment on their rims. Interstitial to this crystalline network is found up to 25% mesostasis that is granitic in composition and has crystals of ilmenite, phosphates, and other accessory minerals. These rocks are not cataclastic, perhaps because crushed versions would not be distinguishable from breccias.

KREEP has major element compositions that plot near the olivine–pyroxene–plagioclase piercing point (Fig. 9) suggesting that they formed either by partial melting, or from the liquid residue of fractional crystallization. Their enriched lithophile trace-element patterns (Fig. 10) are remarkably similar in samples from all over the Moon. This uniformity in pattern is impossible to generate by different episodes of partial melting in different parts of the Moon, indicating that *KREEP* originated from the residual liquid generated by the crystallization of the cumulates that formed the lunar crust. This residual liquid achieved the same degree of fractionation at all the sampled regions of the Moon.

KREEP basalts are dated at 3.9 Ga by Rb–Sr and Nd–Sm internal isochrons.

Polymict breccias are the most abundant rock suite of the highlands, consisting of a mixture of lithic, mineral, and glass clasts set in a finer-grained matrix. The clast assemblage is polymict so that it must have been derived from more than one source rock. The matrix may be a glass-rich, finer-grained version of the clasts (vitric matrix), or it may be an “equilibrated” igneous-textured crystal aggregate (crystalline matrix–impact melt).

Vitric matrix breccias have a seriate texture: the grain size of the constituents displays a continuum from matrix fragments less than a micrometer to clasts several centimeters across. The boundary between matrix and clast is arbitrarily set at 25 μm . The constituents, both clast and matrix consist of glass, mineral, and lithic fragments. Clasts are enriched in lithic fragments and the matrix is enriched in glass fragments. Clast grain shapes are subrounded to subangular, whereas matrix grain shapes are subangular to angular.

Vitric matrix breccias consist of calcic plagioclase, pyroxenes, olivine, ilmenite, glass, and lithic fragments. Accessory minerals are Fe–Ni metal, troilite, phosphate (apatite and whitlockite), K-feldspar, and spinels. The mineralogy is indicative of a detrital assemblage, there being no systematic patterns for the mineral assemblages and compositions.

Crystalline matrix breccias consist of clasts set in a fine-grained matrix. The clasts are texturally and mineralogically similar to the clasts in vitric matrix breccias. The matrix is strikingly different, consisting only of plagioclase, pyroxene, ilmenite, olivine, and spinel. Glass and lithic fragments are not found. Mineral grains are equilibrated in that all the crystals of each phase in a given breccia have about the same shape (either tabular or equant), size, and chemical composition. The matrix texture is not seriate.

The clast assemblage in the finest-grained crystalline matrix breccias is similar to the clast assemblage in vitric matrix breccias except for the lack of glass clasts. As the matrix grain size increases, the clast assemblage progressively responds by (1) reduction in the total percentages of clasts, (2) reduction in the pyroxene:olivine clasts ratio, and (3) change in clast composition to more calcic plagioclase and more magnesian olivine. These systematic changes suggest that the initial clast assemblage was progressively melted, resulting in a present clast assemblage that is biased towards refractory compositions. This clast digestion results from high proportions of melt in the initial breccia mixture. In some breccias the proportion of melt was so great that almost all clasts were digested and the resulting rock is a “breccia without clasts.”

Chemically, the breccias are mixtures of the preexisting highland igneous rocks with meteorite material. Major-element compositions are dominated by mixtures of *KREEP* and highland cumulates, and do not plot near any cotectic thermal valley (Fig. 9). Lithophile trace-element abundances (Fig. 10) are dominated by the pattern imposed by *KREEP*, which is the only component that is enriched in these elements.

Enriched siderophile trace-element concentrations are due to meteorite material that has been added during impact.

Most polymict breccias have crystallization ages (metamorphism or recrystallization) of 3.84–4.05 Ga. The clustering of breccia ages suggests that most craters in the lunar highlands formed during this interval, and the concept of a late, heavy meteorite bombardment derives from this.

Polymict breccias form as a product of meteorite impact, and their characteristics reflect the more important aspects of this process. The mean impact velocity for meteorites hitting the Moon is 20 km/sec: kinetic energy released is such that the impacting body explodes and a shock wave passes through the target rocks with a velocity of hundreds of km/sec. At the point of impact, peak pressures are hundreds of kilobars and particle velocities are hundreds of m/sec. The shock wave is highly attenuated, so that pressures and velocities decay exponentially with distance from the point of impact. Specific phenomena that result include (1) shock metamorphism of minerals, producing shattering, planar features, and vitrification; (2) heating, in which temperature is proportional to the peak pressure (a small volume will be intensely heated and a large volume slightly heated); (3) melting and volatilization, the most highly heated material will be melted and some material will be boiled (mass of melt is a few percent of all ejected material); (4) cracking of rocks; (5) movement of rocks (cracked material will be accelerated so that a crater is formed); and (6) mixing (movement of material is random and the deposits are well homogenized, even from a heterogenous target).

The breccia facies that form in ejecta deposits are considered in the context of a two-component thermal model; cold fragmental material and hot-to-superheated melt material. These materials are violently mixed in such a way that the two components are evenly disseminated one in the other on a scale of a millimeter. Since the mixture is very fine-grained and intermixed, the cold fragments quench adjacent hot melt in seconds, initiating crystallization, hampering differentiation, and giving the resultant breccia matrix a fine-grained texture. The hot breccia then loses heat to its surroundings like a volcanic flow. The proportion of hot melt in the mixture is a function of distance from the point of impact and the size of the impact. The percentage melt in the mixture determines the equilibrium temperature, which in turn determines the breccia facies formed (e.g., impact melt or vitric matrix), and the classes of petrological

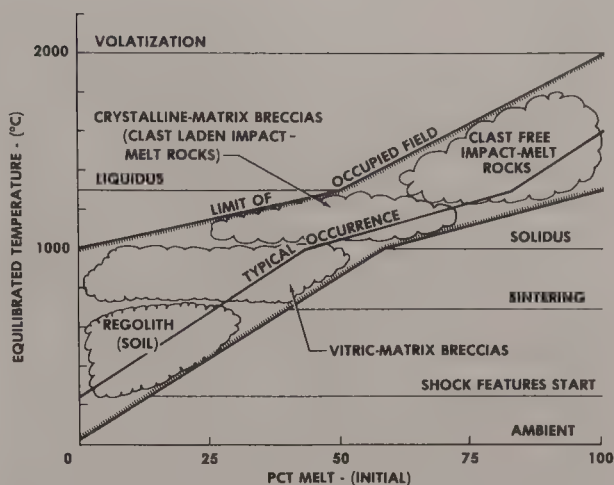


FIGURE 12. Representation of the two-component thermal model for the description of facies in polymict breccias. Distance from the point of impact and impact size determines the percent melt in the ejecta. The percent melt determines the temperature of rapid equilibration (quenching) of melt and included cold clastic material. The breccia facies is determined by the equilibration temperature.

processes that may take place (e.g., clast digestion, rapid nucleation, etc.—Fig. 12). The resulting breccia type is a function of the equilibration. Below the sintering temperature ejecta will not sinter and forms regolith. Between the sintering and solidus temperatures vitric matrix breccias form. Between the solidus and liquidus temperatures clast-laden crystalline matrix breccias result. Above the liquidus temperature clast-poor crystalline matrix breccias form.

Regolith or lunar soil

The lunar surface is covered with a mantle of fragmental debris termed the *regolith*; finer portions of this are commonly, although incorrectly, called soil.

The regolith consists of destructional, constructional, and volcanic fragments, i.e., the results of meteoritic impacts yielding lithic, glass, and individual mineral fragments. Since there is no chemical weathering on the Moon, these fragments are all fresh and unaltered. Comminuted material forms a poorly-sorted, log-normal size-frequency distribution with a mean around 50 μm . Comminuted fragments are angular to subrounded and more or less equant.

No fragment in any lunar soil has been unambiguously identified as being meteoritic, although chemical analysis of siderophile trace elements suggests that up to several percent of most soil samples consist of meteorite material. Fe-metal fragments are probably ultimately derived from meteorites; however, they have

compositions that match metal grains in breccias, so their immediate source is unknown.

Constructional fragments in lunar regolith, termed *agglutinates*, are those particles that have formed during the impact process. They consist of a heterogeneous mixture of comminuted mineral, glass, and lithic fragments, bonded together with vesicular, flow-banded, heterogeneous glass. Agglutinates are generally 1 mm in diameter, and range in abundance from 0 to 60% of lunar soils. They form by the impact of millimeter-sized meteorites which cause melting, and the melt mixes with fragmental material. This process is a miniature version of the breccia-making process.

About 1% of the particles in most lunar soils are of volcanic origin. They consist of glass (clear or devitrified) and occur in regular shapes such as spheres, flattened spheres and dumbbells. Both a soil sample from Apollo 15 (green) and one (orange) from Apollo 17 consist of almost 100% volcanic glass; they are pyroclastic deposits.

Each sample of lunar regolith has had a complex history. It is a mixture of constructional and destructional fragments from an uncountable number of impact events. Some fragments are “new” in the regolith, whereas other fragments have been reworked numerous times. Most fragments are locally derived, whereas a few fragments come from remote locations on the Moon’s surface. Mixing has been particularly efficient in the formation of the regolith. This is evidenced by the fidelity with which the bulk chemical composition of regolith reflects the chemical composition of the local bedrock.

The exposure age of a regolith sample is the average time each particle in the soil has been close to the surface (i.e., within a few centimeters). Exposure age is measured by the accumulation of cosmogenic or solar-wind-derived noble gases (such as ^{38}Ar), light elements (C, N), and high-energy nuclear particle tracks. The content of agglutinates is directly correlated with exposure age and inversely correlated with mean grain size.

A concept of maturity has been developed. Immature soils consist of “new” material (either volcanic or freshly comminuted ejecta). Mature soils consist of a blend of comminuted material and agglutinates (owing to impacts into pre-existing regolith causing continued comminution and agglutinate formation).

The evolution of a soil is thus a combination between maturation and mixing of two (or more) soils. Detailed paths for individual soils become complex. Soils of differing maturity are mixed to form an individual soil: this happened numerous times all over the lunar surface (a process termed “gardening”).

Geological development of the Moon

Although the origin of the Moon is uncertain, a fairly complete understanding of its history can be constructed (Fig. 13). After a 10^8 -year period of nucleosynthesis, isolation of a solar nebula, condensation of mineral grains out of the gas of the nebula, and accretion, the Moon reached its present size 4.55 Ga ago. This period of accretion ended with the outer few hundred kilometers of the Moon constituting a planet-wide “magma-ocean.” Fractionation and crystal

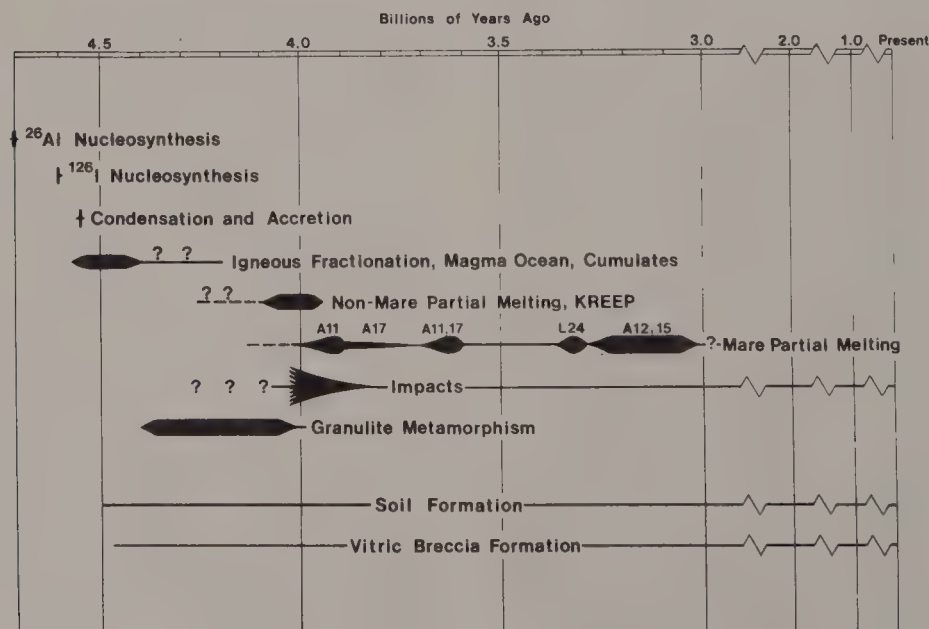


FIGURE 13. Geological history of the moon.

sinking in the magma ocean produced the plutonic rocks that formed an anorthositic crust and a mafic to ultramafic mantle. This was the fractionation process that created the depletion in Eu and Sr that is observed in mare basalts. Late-stage liquid residue from this magma ocean was the parent of the highland volcanic rocks called KREEP. The bulk of the cumulates formed between 4.6 and 4.4 Ga ago and KREEP was extruded at 3.9 Ga, perhaps owing to an effect of the late, heavy meteorite bombardment. The mafic to ultramafic cumulates were subsequently partially melted to form the mare basalts between 3.9 and 3.2 Ga. Photogeological evidence suggests that this period continued until about 2.0 Ga ago.

Contemporaneously with this igneous history, the Moon was being bombarded by meteorites. The time interval of 4.05 to 3.84 Ga represents the age of most of the impacts that produced the craters observed in the highlands, and the bulk of the highland breccias. During this period the great circular basins containing the mare basalts were excavated. At about 3.85 Ga the late heavy bombardment ended and bombardment has continued at a greatly reduced intensity, producing isolated but conspicuous craters like Copernicus and Tycho.

Today, the Moon is inactive except for meteorite impacts that produce the lunar regolith and a few large craters.

J. L. WARNER AND C. H. SIMONDS

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Cross-references: *Anorthosite; Astrobleme; Basalt; Core; Cumulate textures; Flood basalt; Gabbro and norite; Granulite; Intrusion—layered; Magma; Meteorites—petrology; Shock metamorphism; Textures of igneous rocks; Ultramafic rocks; Upper mantle—experimental petrology.*

M

MAFIC AND ULTRAMAFIC INCLUSIONS IN IGNEOUS ROCKS

Mafic and ultramafic inclusions in igneous rocks have for a long time stimulated the interest of geologists and petrologists with peridotite inclusions with ovoidal shape ("olivine nodules") having been studied for more than a century. At first, the interest was the result of the striking contrast of color, mineral assemblage, and texture between the inclusion and host rock, but now a major reason for their study arises from at least some of them being representative of mantle source material for melts of mafic and ultramafic composition. The physical, chemical, and mineralogical properties of these inclusions appear to remain within the geophysical and petrological constraints on the rock types of the upper mantle. Other inclusions consist of minerals and monomineralic or polymineralic rocks (igneous, sedimentary and metamorphic).

Both the presence and abundance of inclusions in igneous rocks are controlled by the extent of stoping of country rock during magma intrusion, and the degree to which stoped material reacts with the magma. In plutonic bodies, inclusions are mostly found close to the margins or roof, where rapid cooling prevented assimilation. Because of the fast cooling rate, those igneous bodies that resulted from rapid transport and emplacement of the magma, such as breccia pipes and dykes, may instead contain a large and varied population of inclusions.

More than 200 occurrences have been reported for peridotite in alkali basalt alone (in flows, pyroclastic deposits and intrusions). There are also numerous occurrences in kimberlite diatremes, alkali ultramafic (and mafic) rocks and other igneous rocks (see Wyllie, 1967; Hawkesworth and Norry, 1983; Menzies, 1987; and Nixon, 1987 for reviews).

In alkali basalts

Unlike mafic inclusions, ultramafic inclusions are everywhere preferentially associated with alkali basalts. This preference is well illustrated in the Hawaiian islands (Kuno, 1969) which consist predominantly of tholeiitic basalt shield volcanoes, capped by younger alkaline lavas

(mugearite, hawaiiite, alkali basalt, and minor trachyte). Except on Hawaii itself, the most recent volcanic activity is represented by flows and pipes of nephelinite, melilite basalt and basanite. The distribution of inclusions amongst these lava types is summarized in Table 1. Ultramafic inclusions (peridotites and pyroxenites) are most numerous in the undersaturated alkaline lavas. Similar associations are found on Kerguelen Island, Reunion, the Galapagos Islands and the Canary Islands, and at most of the world's alkali basalt occurrences (e.g., Midland Valley of Scotland—Upton, 1984).

In kimberlite

Inclusions in kimberlite can be grouped into (1) fragments of adjacent wall-rock, (2) fragments of higher (younger) formations now eroded away (backfall), (3) blocks from buried formations, (4) granulite from the deep crust, and (5) mantle xenoliths. Those of the last group include mainly garnet peridotites and garnet pyroxenites, along with less abundant eclogites, grosspydite (grossular-pyroxene-kyanite eclogite), amphibole and mica peridotites, and glimmerites. The actual suite present at a single locality is highly variable: each province has its peculiarities; e.g., the grosspydites are particularly common in the Yakutia kimberlites of E Siberia. In a single pipe, inclusions are more abundant in the so-called "breccia-facies" than in the massive kimberlite, or may be absent in the pipe, but common in any associated pyroclastic deposits. Fast cooling favors preservation. This also holds true for diamonds, which occur in the xenoliths in some kimberlites and related lamproites (Boyd and Meyer, 1979; Dawson, 1980; Deines et al., 1984).

In alkali ultramafic rocks

Ultramafic inclusions occur in alkalic ultrabasic rocks of ring complexes, in alkalic ultrabasic lavas, and as ejectamenta in associated pyroclastic deposits. Ultrabasic rocks of this type are usually restricted to intracratonic environments, and are associated with alkalic volcanic fields. The occurrence of pyroxenites appears to be restricted to fractured and rifted areas such as those of the Kola Peninsula and

TABLE 1. Type and distribution of inclusions in the lavas of the Hawaiian islands

Zones of the shields	Types of lava	Abundance of inclusions	Types of inclusions
Surficial (youngest)	Nephelinites Basanites Melilite basalts	Very numerous	Peridotites 4-component garnet pyroxenites
Intermediate	Hawaiites Mugearites Alkali basalts	Numerous	Olivine-augite peridotites Dunites Gabbros
Deep (older shield)	Tholeiitic basalts	Less numerous	Gabbros

Baikal (U.S.S.R.) and E and S Africa. The Afrikanda complex in the Kola Peninsula consists of several types of alkalic ultrabasic rock, concentrically distributed, and ranging from pyroxenites and melteigites in the outer portion to coarse pyroxenite in the core. At the Napak volcano in Uganda, xenoliths of pyroxenite occur in a matrix of melteigite-urtite rocks situated in the central part of the volcano, now exposed by erosion. Similar xenoliths occur in associated pyroclastic deposits.

In other igneous rocks

Mafic inclusions, both autolithic and xenolithic, are found in many types of igneous rocks, both intrusive and extrusive, mafic to felsic, in all magma series. In lavas such inclusions may represent (1) fragments from older parts of the volcanic edifice (cognate inclusions), (2) cumulates that are cogenetic with the host lava (autoliths), or (3) plutonic rocks buried at depth, having little or no relationship with the host rock (xenoliths). At Lanzarote in the Canary Islands, cumulate gabbro inclusions occur in alkali basalt, and although basaltic in nature they are unrelated to the host flows. Similar associations occur in other ocean islands. All these cases represent inclusions of cumulates derived from tholeiitic basaltic magma in alkali basalt hosts. Other mafic autoliths, occurring in alkali basalt flows are of cumulate material derived from alkali basalt magma (Carmichael et al., 1974).

Petrogenetic implications

Ultramafic inclusions in igneous rocks may provide information on the source rock from which magmas originate, information on the processes operative during ascent and emplacement, and information about the distribution of rocks encountered along the conduit and entrained by the ascending melt. Ultramafic xenoliths often exhibit a mineral assemblage stable at mantle pressures and temperatures and so provide clues to the origin of magma in the mantle, and to the composition of the upper

mantle itself. To be able to determine which xenoliths, or suites of xenoliths, best represent mantle material, constraints other than those provided by phase equilibria experiments must be considered.

Geophysical measurements of heat flow in oceanic and continental areas have provided an estimate of the temperature distribution at depth (geothermal gradient or geotherm). Data from astronomy and seismology have provided an estimate of the density and pressure distribution within the Earth. Seismic data indicate that the mantle, although solid throughout, may contain molten material in a broad zone (low-velocity zone) at depths of 60–250 km.

When the geotherm for a region is known, the xenoliths of a suite recovered from an igneous body, like a kimberlite pipe, can be plotted on a temperature–depth diagram with the appropriate geothermal gradient; the depth at which the xenoliths equilibrated can thus be determined. When both the pressure and the temperature of equilibration of a xenolith can be determined (geothermobarometry), the curve described by plotting the xenoliths in a pressure–temperature–depth diagram should coincide with a geotherm. This may differ from the modern geotherm, and when combined with radiometric age dating, such studies permit an assessment of the way geotherms evolve with time. When these are combined with trace element and isotopic studies, an assessment can be made of spatial and secular compositional inhomogeneities in the mantle (Menzies, in Hawkesworth and Norry, 1983; Boyd, 1984).

In petrology, the result of experiments in appropriate subsolidus and crystal–melt equilibria, especially those conducted at high pressures, have provided constraints on the types of rock that can occur at differing depths within the mantle. The amount of partial melting that is necessary to provide magma of a given composition from these model source rocks has been evaluated, and the pressures and temperatures at which such melting can occur have been assessed.

In geochemistry, the abundance and distribu-

tion of compatible, incompatible and refractory lithophile elements have been determined in selected xenoliths and melts. Since chondrite (ordinary chondrite or carbonaceous C1) has been considered by many workers to be the representative of the Earth's primordial material, comparison of the distribution of elements in xenoliths and melts to that in chondrite standards is of paramount importance. Such studies provide a test for the chondrite model, and a justification for the selection of some xenoliths as representative of the primitive mantle material.

The rare-earth elements (REE), a geochemically coherent group of elements, have received a great amount of attention in geochemical modeling, especially in regard to the metasomatic processes operative in the upper mantle. These are considered responsible for some discrepancies in the distribution of REE in xenoliths and chondrites (Wilshire, 1984). Distribution coefficients for K, Rb, Sr, Ba, La, Eu, and Ni for liquids in equilibrium at different temperatures with the minerals of peridotites have been estimated to enable the reconstruction of trace-element patterns and the mineralogy of the source rock for basaltic magma. The abundance of strongly incompatible elements such as K/Rb, Rb/Sr, and Ba/Sr, have been used to determine the fraction of the source rock that has to be melted before source and melt can achieve the same ratios. Different ratios could be explained by different degrees of partial melting of the source, or by melting at different depths.

Some selected isotopic systems, namely $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, $^{208}\text{Pb}/^{204}\text{Pb}$, $^{87}\text{Sr}/^{86}\text{Sr}$, and $^{144}\text{Nd}/^{143}\text{Nd}$, have been used to determine the time-related relationship between host rock and xenoliths. One particular relationship is that regarding the age of a piece of mantle, the timing of melt generation and the timing of its eruption.

A recent study of carbon isotope composition in diamond-bearing inclusions from South African kimberlites suggests that a relationship exists between the isotopic composition of the diamonds and the chemical composition of the silicates. Deines et al. (1984) suggest that this indicates the existence of discrete mantle regions characterized by different mean carbon isotope compositions. Studies of Nd isotope systematics of silicate inclusions in diamonds suggest the existence of discrete regions in the mantle that separated from the main mantle reservoir as early as Archaean times and were sampled by kimberlites erupted in the late Mesozoic (Richardson et al., 1984).

Among all the materials selected as representing mantle (i.e., meteorites, basalts, alpine

peridotites, and ophiolites), ultramafic xenoliths offer the most information with regard to (1) the abundance and the distribution of rock-types in the upper mantle, (2) the origin of basaltic and other magmas, and of kimberlite, (3) the nature of metasomatism in the upper mantle, and (4) the evolution of the Earth's crust and upper mantle.

ANNA T. GAVASCI

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Cross-references: *Basalt; Diatreme; Eclogite; Kimberlite; Peridotite; Pyroxenite; Upper mantle—experimental petrology; Xenolith.*

MAGMA

Magma can be defined as high-temperature molten rock. Its familiar extruded form is

volcanic lava, but the term "magma" covers the physical melt from the point of its generation deep in the Earth to the point of its solidification either within or at the surface of the Earth. Magmas may contain substantial proportions of dissolved volatiles (H_2O , CO_2 , etc.—Wyllie, 1979) that are lost, often spectacularly, during volcanic eruptions, but that play an important role in the generation and subsequent differentiation and behavior of the magma. They may also contain suspended crystals that have precipitated out of, or settled into, the magma, or that are remnants of its source rock. For practical purposes, the term magma is also used in a more specific manner, particularly when discussing the petrogenesis and chemistry of igneous rocks, to refer essentially to the liquid silicate portion, free from volatiles and suspended crystals. The reasoning behind this restriction is that (1) petrologists are invariably dealing with the solidified magma as a rock that has lost most of its volatiles, which are therefore an unknown quantity, and (2) crystals can be added to (or lost from) the magma through crystal settling before solidification. Fine-grained volcanic rocks and chilled marginal facies of intrusive rocks are usually taken as representing true magma compositions. Porphyritic igneous rocks may or may not deviate from the composition of the initial magmas. Some coarse-grained igneous rocks represent crystal cumulates with compositions differing markedly from that of the parent magma.

Magma series

The chemical and mineralogical compositions of igneous rocks vary over an extremely wide range. The range in chemistry of rocks representing specific magmas is less wide, but systematic variations do occur and are clearly shown up in chemical variation diagrams. Several types of serial variation have been distinguished and have been named *magma series* or *igneous rock series* (Kuno, 1968). The variations are continuous and characteristic in each series, but the different magma series may be linked by gradational types.

In the first half of this century, when igneous petrology was totally dominated by the principle of gravity-controlled crystal differentiation (Bowen, 1928), the concept of *parental magmas* evolved, with each of the main magma series being derived through repeated differentiation from its own parental, usually basaltic, magma. Although these ideas have been modified to some extent by experimental work, the existence of magma series with distinct chemical and mineralogical characteristics, linked to

particular tectonic or geothermal environments, is universally acknowledged, e.g., the tholeiitic series is found both in oceanic and continental areas, whereas the strongly potassic alkaline series is almost exclusively continental.

Physical properties

In addition to some of the more important properties of magmas discussed below, data on other properties (e.g., electrical and thermal conductivity, thermal expansion, surface tension, ultrasonic wave velocities, etc.) are given by Murase and McBirney (1973) while Maaløe (1985) deals with magma dynamics.

Temperature. There are problems in measuring the temperatures of erupting lavas and reliable values are rare, particularly for the more violent eruptions. Basaltic magmas from Hawaii reach the surface at temperatures of 1,050–1,190°C, while the less-basic andesitic lavas from Etna are erupted at 900–1,050°C. Some reported values for dacites and rhyolites are as low as 750°C. While most basic lavas are completely immobile at this temperature, perceptible movement has been recorded for some basaltic flows. At the mid-oceanic ridges, basalts are erupted at temperatures in the 1,220–1,370°C range.

Experimental work on natural and synthetic rock systems has established the crystallization ranges of most common magma types, with and without volatile components, e.g., at high $P_{\text{H}_2\text{O}}$ granitic magmas can exist at temperatures below 700°C and certain carbonatite magmas can exist at temperatures as low as 590°C at high $P_{\text{H}_2\text{O}+\text{CO}_2}$. Much higher temperatures are recorded for the basalt lava in accessible lava lakes in Hawaii, with liquidus temperatures at, or above 1,200°C, and solidus temperatures near 980°C (Middlemost, 1985). Even higher temperatures (>1,400°C) have been suggested for the ultramafic lavas from which komatiites crystallized.

Density. Magma densities at high temperatures are important in estimating the rates of crystal settling, whether minerals will sink or float, and how far magmas will rise in the crust. They are more critically dependent on composition than on temperature and decrease significantly with the addition of dissolved water. "Dry" basaltic magmas have densities of 2.60–2.78 g/cm³, andesitic magmas 2.4–2.5, and granitic magmas 2.2–2.3. For comparison, the range of density of plagioclase at 1,200°C is 2.53 (An_0) to 2.71 (An_{100}), while densities of olivine tholeiitic basaltic glasses that congealed in the pressure range 0–30 kb are 2.85–3.08 g/cm³ (cf., Kushiro, 1980).

Viscosity. The constant of proportionality, η

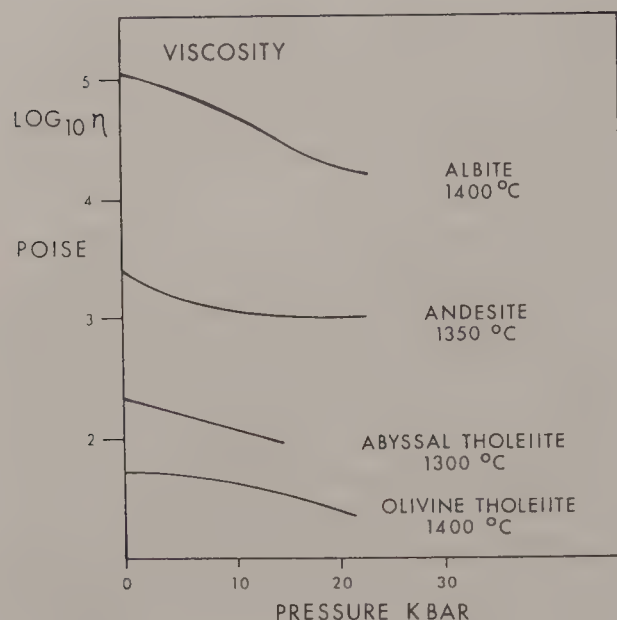


FIGURE 1. Variation in viscosity with pressure for some magmatic compositions and albite liquid. (After Maaløe, 1985.)

(g/cm.s), is the viscosity, which in magmatic liquids decreases with increasing temperature. In dry melts it also decreases with increasing pressure (Fig. 1—Scarfe et al., 1979), a change

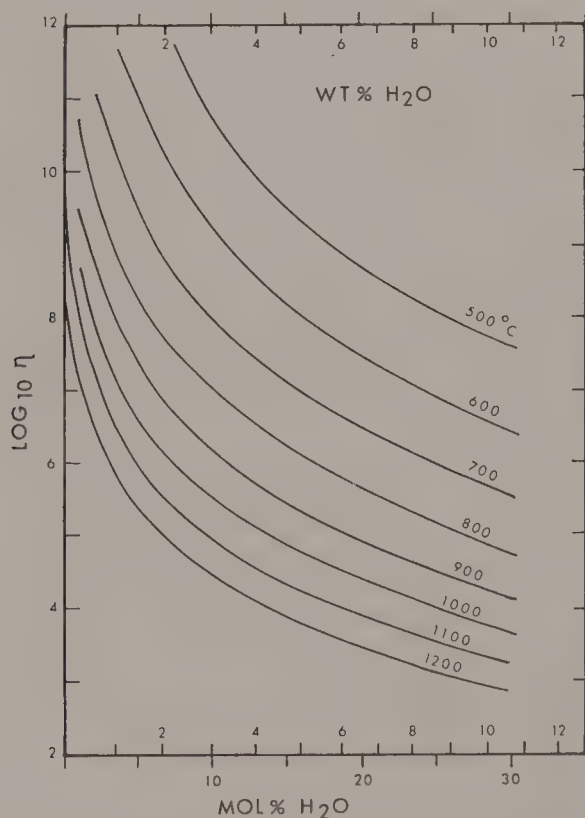


FIGURE 2. Viscosity vs. H_2O content for granitic liquids having bulk compositions approximating to those in the low melting regions of the system Ab-Or-SiO₂-H₂O. The effect of water is greatest for water contents of <3wt.% H_2O . (After Maaløe, 1985.)

related to a change in the polymerization of the silicate liquids. The increased solubility of water with increased pressure in magmas explains the decrease of viscosity in wet melts. This is particularly marked in acidic magmas (Fig. 2) but becomes progressively less important in the more basic magmas.

Experimentally determined viscosities can be related systematically to composition and temperature using parameters such as the glassformer ratio ($O/Si + Al + P$) and the $Al/Al + Si$ ratio, which quantitatively represent the degree of polymerization of the aluminosilicate network in the melt. Similar but more empirical relationships (Bottinga and Weill, 1972) for multicomponent systems enable viscosities to be calculated for a wide range of magma compositions and temperatures. After allowance is made for volatiles, these calculated values agree reasonably with field and laboratory measurements. Basaltic magmas erupting at 1,100°C have viscosities of 2,000–3,000 poises, while for andesitic lavas at ca. 1,000°C the range is 50,000–1,000,000 poises. Measured values for dacitic lavas are as high as 1×10^{10} . Peridotitic magmas, however, would have viscosities <100 poises at temperatures just above their liquidus.

Volatile content. Volatiles such as H_2O and CO_2 have been a major influence on magma generation and magmatic evolution throughout geologic time. In addition, characteristic types of volcanic eruption reflect the volatile content of the erupting lava as much as its composition and viscosity, although these are invariably linked. This contrasts markedly with lunar volcanism, where volatiles have apparently played a minor role.

The estimated volatile content of basaltic magmas during the initial stages of eruption rarely exceeds 1–2 wt.%, decreasing progressively during the eruptive cycle. The dominant gas is usually H_2O (often >90% by volume) with CO_2 the most common gas amongst the remainder, which may include CO , H_2 , HCl , SO_2 , SO_3 , H_2S , and, more rarely, NH_3 , HF , and CH_4 . The reported concentrations of these gases vary considerably between the eruptive and fumarolic stages and between different volcanoes. Air contamination is shown by variable amounts of N_2 , O_2 , and Ar .

The highly significant role in magma behavior played by water has meant that its effects have been studied far more extensively than those of other volatiles (see **Water in rock melts**). However, the role of CO_2 is important in the genesis of highly alkaline igneous rocks, carbonatites, and kimberlites (Bailey, 1980). CO_2 -rich fluids can transport large amounts of Zr, Hf, Nb, Ta, light rare-earth elements, and

P, and there is much evidence for CO_2 playing an important role in metasomatism of subcontinental lithosphere and hence of source regions of magmas (Bailey et al., 1980).

J. TARNEY

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Cross-references: *Alkaline rocks—undersaturated; Andesite and dacite; Archaean volcanism; Basalt; Carbonatite; Explosive volcanic eruptions—classification; Fumarole; Kimberlite; Magmatic differentiation; Ultramafic lavas; Volatiles; Volcanoes and volcanism; Water in rock melts.*

MAGMATIC CRYSTALLIZATION AND PARAGENESIS

Magmatic solutions

A typical magma is probably heterogeneous, with several solid phases, a gaseous phase, and at times a second immiscible liquid phase forming the total mixture. Magmatic liquids are complex, multicomponent solutions, with possible compositions spanning a range as large as that shown by igneous rocks, larger if lost components such as volatiles are considered. The physical and thermodynamic properties of

magmatic solutions are not well known, although accumulating data presents a general picture of the magnitudes of some parameters. For this reason, crystallization is usually discussed by considering the minerals that form, as observed in natural rocks and under controlled experimental conditions. The parageneses (sequences of crystallization) observed can then be compared with mineral, liquid, and whole-rock chemical compositions to describe the crystallization process.

Crystallization interval

Because they are complex solutions, the transition of magmas from liquid to a crystalline solid mixture occurs over a temperature and/or pressure interval. The "liquidus" defines liquid compositions that can coexist with the crystallizing phases during crystallization. Liquidus temperatures are pressure and composition dependent, and describe a surface in T - P - X space, or curves on T - P diagrams at constant composition. The "solidus" is similar to the liquidus, but defines those conditions below which liquid no longer exists and crystallization is complete. Liquidus temperatures are sensitive to relatively small magma composition differences and are less easy to generalize. Solidus temperatures are less sensitive to composition and can be used to compare the conditions at which different magma types exist (Fig. 1).

Temperatures and pressures of crystallization. The temperature and pressure ranges over which magmatic crystallization takes place vary widely; they depend upon magma composition, volatiles, and the position of the magma within the mantle and crust. The lower

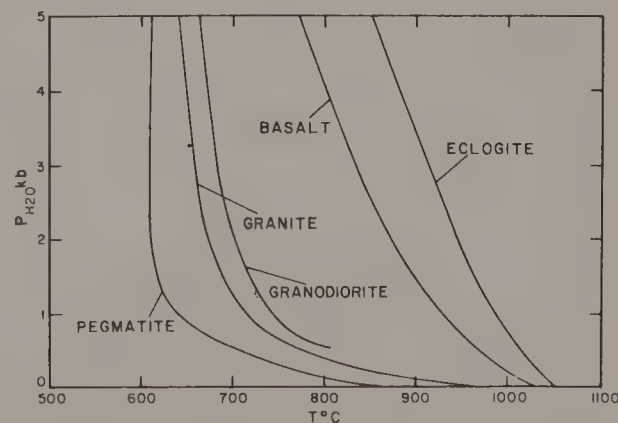


FIGURE 1. Solidus curves for several rock types in the presence of excess water ($P_{\text{TOTAL}} = P_{\text{H}_2\text{O}}$); curves represent conditions at which melting begins or final crystallization occurs; the total crystallization interval for each rock type extends above the solidus to the liquidus for each, respectively; the interval for a granodiorite is shown in Fig. 2.

limit of pressure is restricted only by near-surface conditions where cooling is too rapid for crystallization. However, nearly holocrystalline rocks are found even in thick lava flows. Temperatures have been measured in erupting lavas from 1,200°C to as low as 700°C. Remelting and crystallization of basaltic rocks in the laboratory, at 1 atm. pressure, indicates a crystallization interval of ca. 1,240–1,050°C; granitic compositions are completely crystalline near 960°C. However, at this pressure rocks have lost much of their volatile phases, the presence of which lowers crystallization temperatures.

Influence of volatiles. At higher pressures in the absence of volatiles, the temperature range of the crystallization interval will increase; this can be generally described by application of the Clapeyron equation

$$\frac{dP}{dT} = \frac{\Delta H}{T \Delta V}$$

where P is pressure, T is temperature, H is heat of enthalpy, and V is molar volume. Both the heat and volume terms for crystallization are negative; the solidus and liquidus will have positive slopes on T – P diagrams.

The influence of volatiles is shown by experimental research at high $P_{\text{H}_2\text{O}}$, where crystallization intervals can be lowered several hundred degrees. The magnitude of temperature depression, and the degree to which it continues as pressure increases, depends upon the magma composition. Figure 1 shows that the solidus of granitic magmas is lowered more than 200°C as $P_{\text{H}_2\text{O}}$ increases to 2 kb; further pressure increase has a more limited effect. The depression is related to the large decrease in vapor volume and large increase in water solubility in the magma. Basaltic magma temperatures are less strongly depressed in the lower pressure range, but are depressed more strongly at higher pressures than for granitic magmas. Therefore, although the granite solidus is nearly 300°C below that of basalt at 1 kb, basalt solidus temperatures approach those for granite in excess of 10 kb.

Not all volatiles influence crystallization temperatures the same way. CO_2 has little effect at pressures below ca. 10 kb; the system behaves essentially as if no volatiles were present. Halogen gases (especially F) add to the effect of H_2O to depress temperatures even further. Crystallization temperatures will thus depend upon the presence of a volatile phase and the relative proportions of volatile species, as expressed by their partial pressures. This, in turn, depends upon the solubility of each volatile in the magma and its bulk amount relative to the amount of magma. Solubilities

are temperature-, pressure-, and composition-dependent, indicating that the volatile phase composition will vary during magmatic crystallization. In volatile-deficient magmas, saturation may not occur with respect to one or more volatiles until crystallization has depleted the liquid to a marked degree. Both vapor-absent and vapor-present characteristics will then be displayed through the crystallization interval.

T – P path and saturation. Temperature decrease at constant pressure will cause crystallization. In the absence of volatiles, a pressure decrease will favor melting because of the positive slope of the liquidus and solidus curves and, for crystallization, temperature must decrease sufficiently to overcome this effect. When the pressure is exerted by a volatile such as H_2O (Fig. 1), pressure reduction causes crystallization. It has also been shown that crystallization can take place without the loss of heat (adiabatically) during a pressure reduction; the system will then experience a temperature increase. Such special cases have limited application to nature. A simultaneous decrease in temperature and pressure is probably most common; in some cases pressure increase with temperature decrease may occur.

Magmatic crystallization and mineral paragenesis are essentially problems of mineral saturation in the magmatic solution. Temperature and pressure (dry or volatile) determine when each mineral achieves saturation, but crystallization is largely dependent upon the heat budget and is accompanied by a net loss of heat. The rate of crystallization, or heat loss, is controlled by the rate at which heat can be removed through conduction in the magma and surrounding rock, perhaps aided by processes such as magma convection and gaseous heat transfer. Other factors in the heat budget are the heats of crystallization of the minerals, changing heat capacities of liquid and solid phases, and heats of solution of the crystallizing chemical components in the liquid.

Pressure is important in determining the crystallized rock mineralogy and paragenesis. For example, at pressures >15 kb, a volatile-free basaltic magma will crystallize to eclogite, a rock composed of dense garnet and pyroxene phases. At lower pressures, a less dense basaltic mineralogy will result. If pressure is exerted by H_2O at moderate pressures, the rock will crystallize to amphibolite (quench amphibolite). Common minerals, such as amphiboles and micas, require minimum levels of $P_{\text{H}_2\text{O}}$ for their crystallization and they cannot form if H_2O is absent. Other factors, such as P_{O_2} and Fe–Mg ratios, determine the relative stabilities of minerals containing these components and the temperatures at which they crystallize.

Paragenesis

Once temperature and pressure change so as to cause saturation of the magmatic solution with a mineral phase, crystallization of that mineral can begin. This is expressed on phase diagrams as a phase boundary showing the combinations of temperature, pressure, and liquid composition at which crystal-liquid equilibria are achieved. Some workers have considered the energy problem of nucleating a new phase and suggest that a certain degree of supercooling must take place for the initial crystals to form. This would then be followed by a return of the system to the phase boundary.

Much has been learned about magmatic crystallization from experiments on simple systems in which important rock-forming minerals crystallize. To apply these studies, the lower crystallization temperatures of more complex solutions must be considered, as well as the more variable mineral compositions and additional minerals that will precipitate. Two end-member types of chemical systems are involved. Minerals that are immiscible, or have limited miscibility with other minerals, follow the general rules of eutectic systems. Minerals that can form complete, or extensive, solid solutions with other minerals follow the rules of solid-solution phase diagrams, when a continual change in mineral composition occurs toward compositions more stable at the changed temperatures and pressures. Both types of crystallization occur in magmas, individually or simultaneously, and there are gradations between the two. Many common minerals display extensive solid solution, including the olivines, pyroxenes, plagioclases, alkali feldspars, amphiboles, and micas. The ranges of compositions formed in a magma depend upon magma composition and the temperature-pressure range of crystallization. The allowed extent of solid solution may decrease at lower temperatures, as in the alkali feldspars. High-temperature alkali feldspar solutions can unmix at low temperatures to form perthitic intergrowths. Simultaneous crystallization of nearly pure alkali feldspar and quartz at low temperatures will closely follow the characteristics of a eutectic system.

Granodiorite-water system paragenesis

Experimental research on natural rocks simulates the compositional complexity of natural magmas. Crystallization of volatile-absent rocks usually is limited to studies at atmospheric pressure, because of the high temperatures required at higher pressures and slow reaction rates. Extensive work on rocks in the presence of excess volatiles, especially H_2O ,

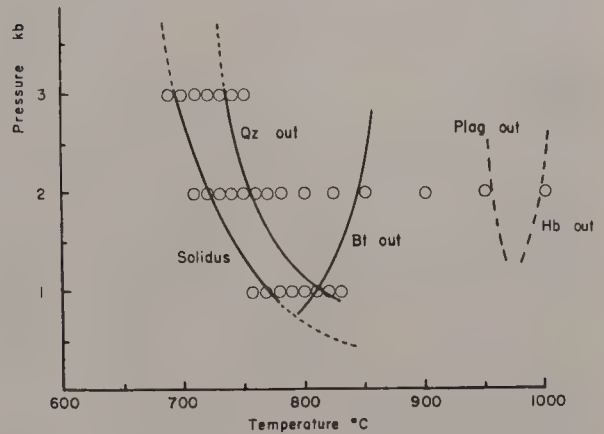


FIGURE 2. Results of melting experiments on a granodiorite in the presence of excess water ($P_{TOTAL} = P_{H_2O}$); curves show conditions at which each mineral phase became completely melted and they can be interpreted as also showing approximate conditions for initial crystallization of each phase; circles indicate conditions of various experiments. Hb, hornblende; Plag, plagioclase; Bt, biotite; Qz, quartz; Kspar, alkali feldspar. (From Piwinski and Wyllie, 1968.)

shows parageneses that are in general agreement with field and petrographic data. Figure 2 shows an example of experimental melting of a granodiorite with excess water ($P_{TOTAL} = P_{H_2O}$). The curves show where melting begins (solidus) and where each mineral phase becomes completely melted. The diagram can be considered to show the general features of crystallization since, at equilibrium, the reverse process of initial crystallization of each mineral would be expected near the same curves. At constant pressure of about 2 kb, a temperature decrease to near 950°C will cause crystallization of hornblende to begin. If this is the first mineral to crystallize, this curve also defines the maximum liquidus temperature for the system. As the temperature decreases below 950°C, hornblende crystals grow until the plagioclase curve is reached. At this temperature (about 915°C) plagioclase begins to crystallize simultaneously with hornblende; a phase boundary for the coexistence of hornblende-plagioclase-liquid is now followed. The process continues with biotite, then quartz, and finally alkali feldspar joining the previous minerals. All alkali feldspar crystallization takes place within a small temperature interval just above the solidus. The proportions of minerals present after crystallization is complete, and their order of appearance (paragenesis), is dependent upon the initial magma composition. The same minerals can form, in different amounts with a different sequence, from a magma of considerably different bulk composition. Also, a rock more deficient in alkalies, such as a tonalite, may be depleted in alkalies during

amphibole and biotite crystallization, preventing the formation of a separate alkali feldspar phase.

If each mineral remains stable through the crystallization interval, it may continue to precipitate until the solidus is reached; however, the temperature interval over which most of its crystallization occurs can be small. In the system shown (Fig. 2) nearly all hornblende and plagioclase are precipitated above 900°C at 2 kb. Since they are solid solutions, these minerals will tend to react with the liquid as temperature decreases. If heat removal is rapid, mineral zoning may occur, with the outer zones successively indicating the compositions more stable at lower temperatures. In Fig. 2, crystallization is controlled by solid-solution phase relationships at higher temperatures. Eutectic phase relationships are more important during the period dominated by quartz and alkali feldspar coprecipitation. Hornblende and plagioclase phenocrysts and "eutectic textures" between quartz and alkali feldspars in many natural granodiorites corroborate the basic features suggested by this experimental system.

Effects of liquid depletion

The amount of liquid decreases the crystallization interval; in general, its proportion of the total mass will not vary linearly with temperature. Figure 3 shows the approximate percentages of liquid and crystals at various stages for the system of Fig. 2 and indicates two "pulses" of crystallization. Above 900°C, the liquid fraction decreases to nearly 50%; from 900°C to

ca. 730°C little further depletion occurs; below 730°C all of the remaining liquid is consumed. This suggests that the rate at which temperature decreases may vary. If the rate of heat removal is restricted, temperature decrease during crystallization "pulses" will be slower because of the additional heat of crystallization that must be removed. Temperature decrease across the "plateau" would be more rapid. Conversely, if heat removal were efficient, very rapid crystallization would be expected during the two pulses of crystallization. Several other results of nonlinear liquid depletion can be expected.

Liquid depletion will control volatile-phase compositions and the point at which a bubble phase appears. With excess volatiles, the liquid is always saturated with volatiles and a vapor phase exists throughout. Volatile-deficient systems may not develop a volatile phase until late in the crystallization history, when the amount of volatiles is sufficient to cause saturation. In some cases, the crystallization of hydrous minerals (amphiboles, micas) may significantly deplete the magma in volatiles. Since several volatile species will be present, each with different solubilities and initial amounts, saturation with respect to each can occur at different times. The volatile phase composition will then change as crystallization proceeds.

The magmatic liquid is depleted in precipitating components and its composition continually changes. Early minerals have a controlling influence on when saturation with respect to later minerals is achieved. If various differentiation mechanisms occur, such as separation of early minerals from the liquid, the expected paragenesis of the rock and its total mineralogy will be changed. In Fig. 2, the initial granodioritic magma becomes enriched in K and Si; the liquid fraction attains a composition like that of a simple granite. Separation of liquid from crystals at this time will lead to the formation of a granitic or pegmatitic igneous body, leaving a rock more basic than granodiorite behind.

Liquid compositions across the crystallization interval can be obtained numerically by subtracting the proper proportions of mineral compositions at different stages. Detailed experimental studies, on whole rocks or specific simple systems, give liquid compositions under specified conditions. These are shown on phase diagrams as paths on the liquidus surface. Geometrical constructions using tie lines which connect the crystals coexisting with various liquids along the path will give the proportions of phases present. A major petrological problem is to apply such data accurately to natural systems or obtain this information from the rocks themselves. In volcanic rocks, where

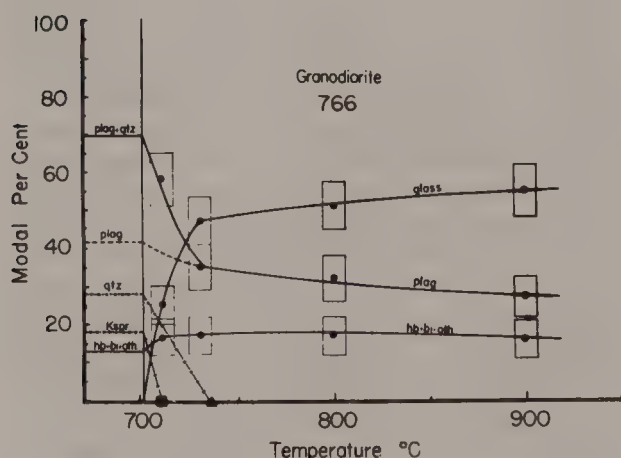


FIGURE 3. Percentages of mineral phases and liquid (glass) present across the melting (crystallization) interval for the granodiorite shown in Fig. 2. Pressure is 2 kb (P_{H_2O}); solidus is at 700°C; values below solidus indicate initial rock composition; liquidus is near 950°C; the position of the "hornblende out" curve, at 2 kb in Fig. 2; (From Piwinski and Wyllie, 1968.)

residual liquids are quenched to glasses, compositions can be determined with some accuracy. In holocrystalline rocks, estimates can be made by petrographically determining the paragenesis and mineral proportions. A series of whole-rock compositions, that appear to be cogenetic, can also be compared to seek compositional trends that may suggest liquid compositions in a parent magma chamber. Such compositions can be shown on liquid descent, or variation diagrams, or many other types of diagrams in which certain major oxides or chemically related oxides are compared.

An important aspect of liquid depletion is the changing physical properties of the remaining liquid. Its changing density, viscosity, heat capacity, and other parameters can influence significantly diffusion, crystal growth rates, differentiation processes, and magma movement. These factors will, in large part, determine the texture and mineralogy of the rock.

BERT E. NORDLIE

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Cross-references: *Amphibolite; Basalt; Cumulate textures; Eclogite; Experimental petrology; Geochemical modeling in petrogenesis; Granodiorite; Liquid immiscibility; Magma; Magmatic differentiation; Nucleation of minerals in magmas; Phase rule; Silicate melting; Textures of igneous rocks; Volatiles.*

MAGMATIC CRYSTALLIZATION IN METAMORPHIC ENVIRONMENTS

The most obvious effect seen in bodies of magma emplaced and crystallized under metamorphic conditions is the increased grain size resulting from a lower than normal cooling rate, as instanced by the gabbroic nature of some basic dykes in deeply eroded Precambrian terrane. This arises because viscosity and diffusion, which control crystal growth in

magmas, are markedly temperature-dependent, while the rate of nucleation of mineral phases is related to the rate of cooling. Another effect can be the enrichment in marginal phases of early-crystallizing mineral phases because the combination of hot country rocks associated with low-viscosity magma keeps T near the margins high enough during magma intrusion to prevent precipitation of the later crystallizing phases (Tarney, 1973).

With high confining pressure preventing the escape of volatiles, magmas that crystallized at depth commonly contain hydrous minerals such as hornblende and biotite. The temperature at which amphibole is the stable crystallizing phase in basaltic magmas is increased considerably as P_{H_2O} increases (Yoder and Tilley, 1962) and many Precambrian dolerite dykes contain appreciable primary hornblende; those consisting almost essentially of amphibole are termed *quench amphibolites*. Coarse-grained "dolerite" dykes with relatively fresh chilled margins but with amphibolite, garnet amphibolite, and foliated garnet amphibolite centers have been recorded from W Greenland (Windley, 1970). These features are attributed to the build-up of volatile components during the later stages of crystallization, with late-stage crystallization taking place under stress.

Continuing high P – T conditions following crystallization may initiate solid-state reactions between the igneous minerals. O'Hara (1961) has attributed the growth of abundant small granular garnets in the fine-grained margins of basic dykes in NW Scotland to autometamorphism: the higher temperatures caused by the dykes superimposed on already hot country rocks initiated reactions that caused the fine-grained margins to reequilibrate to regional conditions. Similar effects are frequently seen in igneous rocks emplaced into high-grade metamorphic terranes, but may pervade the whole igneous body. Reaction rims (coronas) between olivine and plagioclase and between plagioclase and orthopyroxene have been observed in basic intrusions in many parts of the world (Griffin and Heier, 1973, and references therein). Their constituent minerals depend on pressure, temperature, and P_{H_2O} conditions under which the solid-state reactions occurred and include garnet, spinel, pyroxenes, and hornblende. Their environment of formation is that of slow cooling of igneous bodies at moderate to high pressure in metamorphic environments.

Many important secondary postcrystallization effects are seen in igneous rocks that have crystallized in metamorphic environments. A common feature, exhibited by dolerites in high-grade basement terranes and by anorthosites in granulite terranes, is the strong

reddish-brown clouding of their plagioclase. The clouding is caused by exsolution of fine particles of ferric oxide, but there is still dispute as to whether the phenomenon is entirely a secondary effect or whether it is a primary effect accentuated by the prevailing synmetamorphic conditions.

Separation of these late-stage recrystallization effects from the normal prograde effects of regional metamorphism is particularly difficult, especially where high P_{H_2O} and/or deformation have accelerated the reactions. This problem is especially acute with rocks of the anorthosite-mangerite-charnockite suite. These often very large bodies have igneous characteristics yet in many cases appear to have been thoroughly recrystallized and even foliated shortly after they were emplaced, and can quite justifiably be regarded as meta-igneous rocks. Mostly they tend to occur in high-grade granulite facies terranes. However, the extent to which their present metamorphic features have arisen through recrystallization in response to their own enormous heat source and emplacement, rather than through normal high-grade regional metamorphism, is a matter for debate.

J. TARNEY

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Cross-references: *Amphibolite*; *Anorthosite*; *Charnockite suite*; *Metamorphic convergence*; *Nucleation of minerals in magma*; *Reaction rim*; *Water in rock melts*.

MAGMATIC CRYSTALLIZATION IN OPEN SYSTEMS

Crystal fractionation in closed-system magma chambers, involving separation of more mafic cumulate rocks from an increasingly acidic

residual liquid, has long been considered the most important source of variation in suites of igneous rocks. Such systems can be modeled mathematically, and the variations in concentration of each trace element can be predicted from a knowledge of the major-element chemistry and mineral-liquid distribution coefficients for the elements. Closed-system fractionation models explain many features of some igneous suites, especially those that show a tholeiitic differentiation trend. In contrast, no closed-system model can satisfactorily explain the chemistry of any calc-alkaline suite over more than a very restricted range of compositions.

If magma chambers are regarded as open systems, it is possible to explain many features of igneous rocks that cannot be due to closed-system fractionation. Open systems are, however, too complex for exact mathematical modeling at present. They involve too many independent processes, such as replenishment of the magma chamber with fresh magma, loss of fractionated magma by eruption, and assimilation of country rock, all of which open the magmatic system.

Other processes effectively split the magma chamber into open subsystems that evolve internally but are interdependent. For example, double-diffusive convection may split a body of magma into layers of different composition, separated by diffusive interfaces. The full effects of this on the chemistry of the magma are not yet known.

Crystal fractionation splits the chamber into a cumulate pile and residual liquid. Within the cumulate pile, magmatic infiltration metasomatism may occur, with percolating magma reacting with the cumulus material, changing the composition of both (Irvine, 1980). The interstitial liquid may then return to the main body of magma and affect its composition.

Theoretical models of magma chambers as open systems provide qualitative examples of the chemical effects that might be expected, although no such model can encompass all the possible processes in a magma chamber. It is especially difficult to separate the effects of replenishment and assimilation, because both add material of unknown composition to the system. Theoretical models assume that either replenishment (O'Hara and Mathews, 1981) or assimilation (DePaolo, 1981) is dominant, with the other playing only a minor role. These models may be used to illustrate the possible effects of open-system fractionation, but they are too simple to allow the complete mathematical modeling of any natural example.

In the model of DePaolo, fractional crystallization at the base provides energy for

assimilation at the top of a magma chamber, enabling the intrusion to stope its way through the country rocks. Replenishment, eruption, and some assimilation may all occur in the more complex model of O'Hara and Mathews. In both cases, however, an equilibrium situation is eventually reached, if fractionation continues long enough (Figs. 1 and 2).

Many of the essential characteristics of closed-system fractionation do not necessarily apply when the system is open. The composition of a chilled margin is often used to indicate parental magma composition. However, if the magma is periodically replenished, the composition of the batch that formed the margin need not be typical of the intrusion as a whole.

Closed-system fractionation results in a liquid line of descent, a succession of liquid compositions formed by the fractionation of the preceding liquid. This gives rise to trends on variation diagrams. Open-system fractionation also gives rise to such trends, but in these each composition is the result of a balance of igneous processes and the trends observed reflect shifts in this balance.

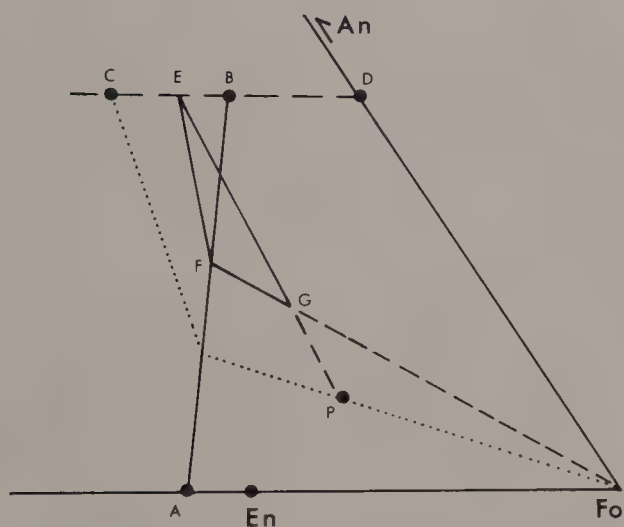


FIGURE 1. Part of the system forsterite-anorthite-silica (schematic), illustrating the fractionation of a parental magma (P) under both closed-system and open-system conditions. In a closed system, P fractionates forsterite (Fo) initially until it reaches the enstatite-forsterite cotectic (A-B), when it starts to fractionate enstatite (En) and moves towards C. Periodic replenishment of the magma chamber will eventually result in an equilibrium condition such as the one in which fractionation of Fo from magma G leads to composition F, followed by En fractionation to E. An influx of P would then return the magma composition to G. If replenishment were continuous, the magma composition would evolve to B and remain there. Variations in the rate of replenishment will result in a scatter of points between P and B, but not forming a straight line between the two. (After O'Hara and Mathews, 1981.)

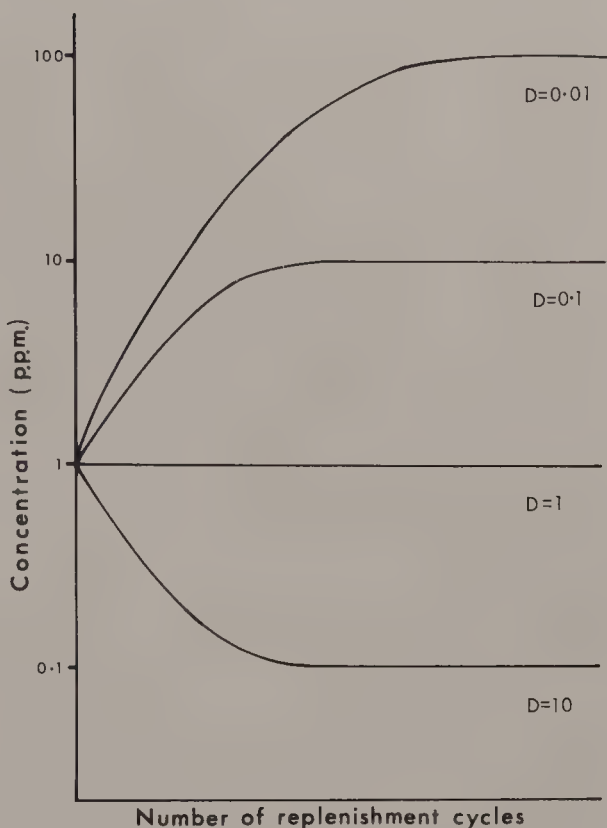


FIGURE 2. Variation of trace-element concentrations with replenishment of a magma chamber (schematic). Replenishment of a fractionating magma chamber eventually leads to an equilibrium concentration for each element that is its initial concentration divided by D , the bulk solid/liquid distribution coefficient for the element; two elements with slightly different values of D will be enriched or depleted to different extents by replenishment of the magma, changing their ratio. (After O'Hara and Mathews, 1981.)

The parental magma composition must lie on the closed-system liquid line of descent, but it need not even be close to trends of variation due to open-system processes (Fig. 1). In an intrusion that formed as a closed system, rocks with cumulate textures cannot represent liquid compositions. If an open-system magma chamber undergoing only replenishment and crystal fractionation were to reach an equilibrium condition, however, the composition of the parental magma being added and of the cumulus material fractionating would be the same. Such an extreme situation is very unlikely to occur, but it illustrates that few constraints can be put on the parental magma composition in an open system.

In a closed system, the only minerals that can fractionate from a magma are those that are on the liquidus and actually crystallizing. This is not necessarily the case in more complex systems. If the magma is undergoing double-

diffusive convection, basaltic liquid at the base might be overlain by granitic liquid. If the basal liquid were to fractionate olivine, the effects of this could be transmitted to the granitic liquid by diffusion. This liquid would thus appear to fractionate a mineral not stable on its liquidus. It should be noted that this granitic liquid would be superheated, yet fractionating, a situation impossible in a simple closed-system magma chamber.

If magmatic infiltration metasomatism occurs, the composition of cumulates may be changed after they have been deposited. For example, it has often been suggested that differentiation in calc-alkaline suites involves fractionation of hornblende and plagioclase. These minerals do not, however, occur together on the liquidus at basaltic compositions, and under all condition where hornblende is on the liquidus so too is another mafic mineral. This means that crystal fractionation of hornblende and plagioclase is not possible. If, however, a calc-alkaline magma initially fractionates olivine and pyroxene, it can then filter down through this cumulus material, reacting with it to produce hornblende. The residual interstitial liquid can then rise back to the main body of magma, fractionating plagioclase as it does so. The net effect is that hornblende and plagioclase are removed from the initial magma composition, with little fractionation of olivine or pyroxene.

Open-system fractionation is a very effective means of "decoupling" trace elements, i.e., of changing the ratios of elements that would be expected to behave in a similar manner (Fig. 2). Theoretical models show that there is a limit to the extent of these changes in a single magma chamber, but where magma ascends through a succession of open-system chambers no such limit would apply. It has been suggested that open-system fractionation is even capable of changing isotopic ratios, but this is not widely accepted.

The trace-element variations expected from open-system fractionation are similar to those expected from partial melting of heterogeneous mantle followed by closed-system fractionation. It is not usually possible to distinguish reliably between these alternatives. In many cases where mantle heterogeneity has been proposed, open-system fractionation has been considered inadequately or not at all.

I. W. FERGUSON

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MAGMATIC CURRENTS AND FLOW

Movement of magma may occur either as currents or as flow. The currents occur in a closed system and are caused by a difference in the density of the magma in different parts of an intrusive body. In an open magmatic system, the flow of magma is caused by pressure differences.

Magmatic currents

The activity of magmatic currents is evident in layered intrusions. A lineated fabric in the accumulated layers and the presence of troughs with a systematic orientation in relation to the structure of the intrusion suggest the activity of large-scale convective currents (Brothers, 1964). Lamination of the accumulated layers may also have been caused by convection, but compaction of these layers may also produce a laminated fabric; lamination of the accumulated layers affords no direct indication of magmatic currents. Lamination and banding of the steep-sided and roofing border facies of the intrusions, on the other hand, indicates current activity.

The physical conditions for the onset of convective currents of a liquid in a sheet-shaped body is given by the dimensionless Rayleigh ratio R :

$$R = \frac{c\beta g d^2 L^3 \Delta T}{\mu k} \geq R_c$$

(where c is the specific heat, β is the coefficient

of thermal expansion, g is the acceleration due to gravity, d is the density, L is the height of sheet, ΔT is the temperature difference between the roof and bottom region of the sheet, μ is the viscosity, and k is the thermal conductivity). For $R < R_c$ the liquid will be in a stationary state, and for $R > R_c$ the liquid will be in a convective state. R_c depends on the actual shape of the body and the boundary conditions but is usually of the order of 1,700. For $R < 10^9$ the convective currents will be laminar, and for $R > 10^9$ the currents will be turbulent (Kay, 1963).

Magmatic flow

Magmatic flow is evidenced from the very presence of intrusions, but flow activity (primary flow) may also be revealed by the internal textural and structural relationships of intrusions. Trachytoid textures and a concentration of prekinetic phenocrysts at the center of dykes indicate flow (Bhattacharji, 1967). A parallel orientation of elongate phenocrysts and xenoliths may indicate the flow direction (Tweto, 1951). Flow structures in granitic batholiths such as schlieren, banding, and trachytoid textures indicate the emplacement direction of the granitic magma (Balk, 1937).

The relation between pressure and mean flow velocity (v_l), by laminar flow between parallel walls is given by

$$v_l = \frac{\Delta P l^2}{12\mu z}$$

where ΔP is pressure difference between the ends of the channel, l is the width of the channel, μ is the viscosity of fluid, and z is the length of the channel (Kay, 1963).

For turbulent flow in a similar channel, the mean velocity (v_t) is given by

$$v_t = 0.0049 N^{3/4} \frac{\Delta P l^2}{12\mu z}$$

where N is the Reynolds number ($N = (dV_m \cdot 2l)/\mu$), d is the density of fluid, v_m is the mean velocity of fluid, l is the width of channel, and μ is the viscosity of fluid (Jacob, 1949). For N less than 2,000 the flow will be laminar; for larger values it will be turbulent.

S. MAALØE

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MAGMATIC DIFFERENTIATION

The evolution of different igneous rocks from a common parental magma is called *magmatic differentiation* (Fig. 1). This term is also used to describe processes that generate heterogeneities in a magma, and then preserve these separate fractions when the magma congeals. It is particularly convenient to use the concept of magmatic differentiation (segregation) when attempting to explain (1) the origin of sequences of volcanic rocks that have systematic differences in composition but have issued from the same volcanic vent, or (2) the systematic variations in composition found in many layered intrusions. Layered intrusions, such as the Skaergaard intrusion of E Greenland, are examples of extreme magmatic differentiation; they show a gradual upward change in composition from minerals that formed at higher temperatures to minerals that precipitated at lower temperatures. The compositions of the plagioclase feldspars, for example, change continuously from being calcic (An_{85}) to being alkalic-calcic (An_{30}). The first tentative ideas on magmatic differentiation were proposed by G. P. Scrope and L. von Buch in 1825. Later, Charles Darwin discussed the concept more fully in a section of his book, *Geological Observations on the Volcanic Islands*, which was

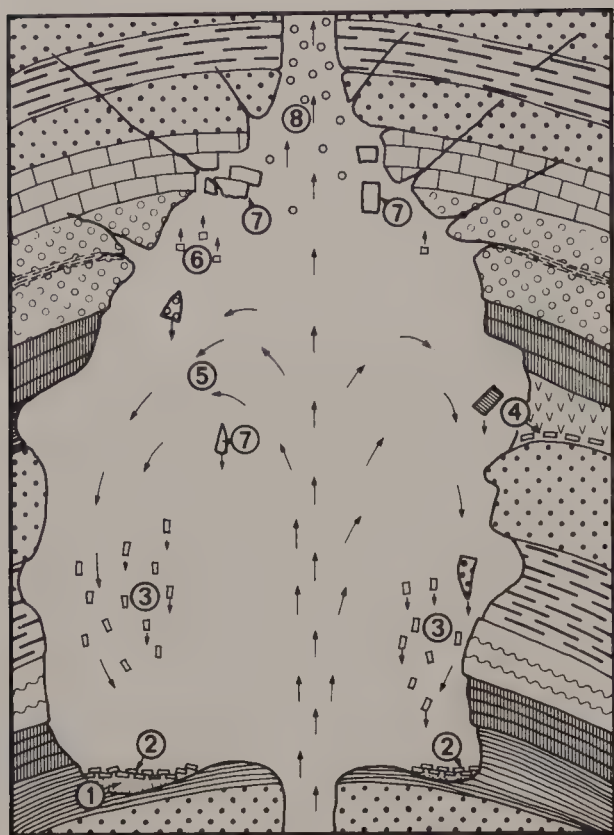


FIGURE 1. Schematic representation of magmatic differentiation in action; 1, sulfides deposited as an immiscible liquid; 2, a mush of Mg-rich olivine crystals; 3, olivine crystals settling under the action of gravity; 4, dyke with an olivine-rich layer at the base; 5, convection currents; 6, upward movement of light sodalite crystals; 7, magmatic stoping and the assimilation of wall rocks; 8, upward movement of volatile constituents such as H_2O and CO_2 .

entitled "Separation of the constituent minerals of lavas, according to their specific gravities."

Liquid immiscibility

The formation of the core and mantle was probably the first major differentiation process within the Earth. This separation occurred because ultramafic liquids and metal-sulfide liquids are immiscible. Liquid immiscibility is an important magmatic differentiation process. It may occur in wholly liquid magmas, or in residual magmas, and it consists of the separation of an initially homogeneous liquid into two compositionally distinct liquid phases. Many early petrologists (see Harker, 1909) used silicate liquid immiscibility in magmas to explain the juxtaposition of rocks such as basalts and rhyolites and the apparent dearth of associated intermediate rocks. Greig (1927) described experimental evidence of immiscibility in anhydrous silicate melts, but these data seemed to show that liquid immiscibility only occurred in melts with compositions that were unlike

those of the common magmatic rocks, and at temperatures ($\sim 1,700^\circ C$) that were excessively high for normal magmas. In more recent times, experimental studies have shown that liquid immiscibility can occur at lower temperatures ($\sim 1,100^\circ$) in a variety of hydrous silicate systems, particularly if these liquids contain minor oxides such as P_2O_5 . The mare basalts of the Moon, and some of the subalkaline basalts of Hawaii, provide clear examples of late-stage liquid immiscibility. In the mesostasis of these rocks there are two coexisting glasses separated by sharp contacts. One glass usually has high SiO_2 and K_2O values, whereas the other glass has low SiO_2 , Al_2O_3 , and K_2O values but high FeO values.

It has also been proposed that the plagiogranites of the ocean crust are derived from an immiscible liquid that evolved late in the crystallization of a magma with a bulk composition similar to MORB (mid-ocean ridge basalt). There is also unambiguous field and laboratory evidence of immiscibility between carbonate-rich fluids and various silicate liquids, particularly alkalic silicate liquids, and it is now generally accepted that immiscibility is an important process in the evolution of the carbonatite-nephelinite rock association (von Eckermann, 1966; Le Bas, 1977). According to Roedder (1979) there is evidence for liquid immiscibility in a wide range of magmatic rocks, and that such immiscibility usually yields a felsic, alkali-aluminosilicate liquid, and a mafic liquid rich in Fe, Mg, Ca, and Ti. There is also field and laboratory evidence of immiscibility between carbonate-rich fluids and various silicate liquids, particularly alkalic silicate liquids (cf. van Groos and Wyllie, 1973). The process also plays a role in the differentiation of many of the alkalic basic and ultrabasic dyke rocks (lamprophyres) associated with carbonatite intrusive complexes (cf., Currie and Ferguson, 1972). According to some petrologists (cf., Jahns and Burnham, 1969) granitic pegmatites may form at a late stage from an immiscible water-rich fluid that separates from a much larger body of hydrous granitic magma.

Soret effect and thermal diffusion-convection

The Soret effect is essentially the tendency of liquids that contain temperature gradients to develop gradients in composition. In the late 19th century this effect was regarded as being one of the main causes of magmatic differentiation, and, when combined with convection, thermal diffusion-convection was considered a significant cause of magmatic differentiation. This process has much in common with a number of other differentiation

processes that have been proposed in more recent times. They include convection-driven thermogravitational diffusion and multidiffusion (see **Double-diffusive convection**).

Another differentiation process that may operate in completely molten magmas is gaseous transfer (pneumatolytic differentiation). Gas moves from lower to higher levels in a body of magma whenever the internal gas pressure is sufficient to produce a separate gas phase. The efficacy of gases as agents of magmatic differentiation lies both in their ability to remove materials when they escape from a magma, and in their ability to selectively scavenge, and transfer, material from the lower to the upper levels of a magma chamber.

The process of convection-driven thermogravitational diffusion was introduced to account for the systematic variations in chemical composition that develop in large bodies of silicic magma. Hildreth (1979) has proposed that the hydrous silicic magma that evolved in the upper part of the magma chamber beneath the Long Valley caldera, California, was generated by the combined effects of convective circulation, internal diffusion, complexing, and wall-rock exchange. The concept of compositionally zoned magma chambers, and the process of convection-driven thermogravitational diffusion that generates them, are both important in developing models that account for the origin of the voluminous silicic ash-flow deposits that occur in many parts of the world.

Multidiffusive convection

In recent years a number of petrologists have attempted to evaluate the nature of flow phenomena in density-stratified magmas. For example, it is known that when hot saline water is released at the bottom of a body of cooler and less saline water, the properties of salinity and temperature have opposing effects on the density of the introduced liquid. High salinity makes the liquid more dense, whereas thermal expansion makes the liquid less dense. During the mixing of the introduced liquid into the main body of liquid, there is a double-diffusive effect, because heat and salt diffuse at different rates. If hot saline water moves upwards through a body of less-saline water that has a stable density gradient, the introduced liquid rises until its density matches that of the surrounding liquid, and it then spreads out horizontally. The introduced liquid is still hot and saline relative to the surrounding liquid, and a diffusive interface develops above it. Heat is exchanged relatively rapidly through the diffusive interface, and that part of the introduced liquid that gives up some of its heat

is now more dense than the liquid below it, and sinks. Such double-diffusive convection is likely to operate in relatively simple chemical systems. In most natural magmas, multidiffusive convection is more likely to occur, with a number of different chemical components being exchanged at different rates through diffusive interfaces (cf., Rice, 1981). Once multidiffusive convection operates in a body of magma, it becomes gravitationally stratified in such a way that it is subdivided into a number of roughly horizontal layers composed of many small convection cells. Each separately convecting layer is likely to become relatively uniform in temperature and composition, and the magma body as a whole is likely to have step-like gradients in temperature and composition. Multidiffusive convection is considered to account for many of the petrographic and geochemical features of layered intrusions (cf., Irvine, 1982).

Fractional crystallization

During the cooling of a magma, crystals usually nucleate and grow. If they react continuously with the magma, the process is called equilibrium crystallization. If the crystals are prevented from reacting and equilibrating with the magma, the process is called fractional crystallization. Both processes are likely to occur during the crystallization of a magma. Fractional crystallization essentially consists of separating minerals, particularly the early-formed minerals, from a residual magma. This is readily accomplished by the mantling, or the mechanical removal, of minerals. In a simple example of the latter processes, minerals precipitate and then sink onto the floor of a magma chamber, where they are covered by other crystals and prevented from reacting with the residual liquid.

In order to understand the nature of the crystallization processes that occur in magmatic rocks, awareness is required of the concept of the reaction series as enunciated by N. L. Bowen. He recognized two different types of reaction series, a continuous reaction series, as exemplified by the plagioclase feldspars, and a discontinuous reaction series that contains the common ferromagnesian minerals (Fig. 2a). A well-studied example of a discontinuous reaction pair is olivine $\rightarrow$ pyroxene in the system $\text{Mg}_2\text{SiO}_4\text{--SiO}_2$. An examination of the reaction series characteristic of subalkaline basaltic magmas crystallizing at low pressures in an essentially closed system (Fig. 2b) reveals that fractional crystallization, with the removal of the early-formed olivines and Mg-pyroxenes and calcic plagioclases, is likely to impoverish the residual liquid in MgO and CaO and enrich it in

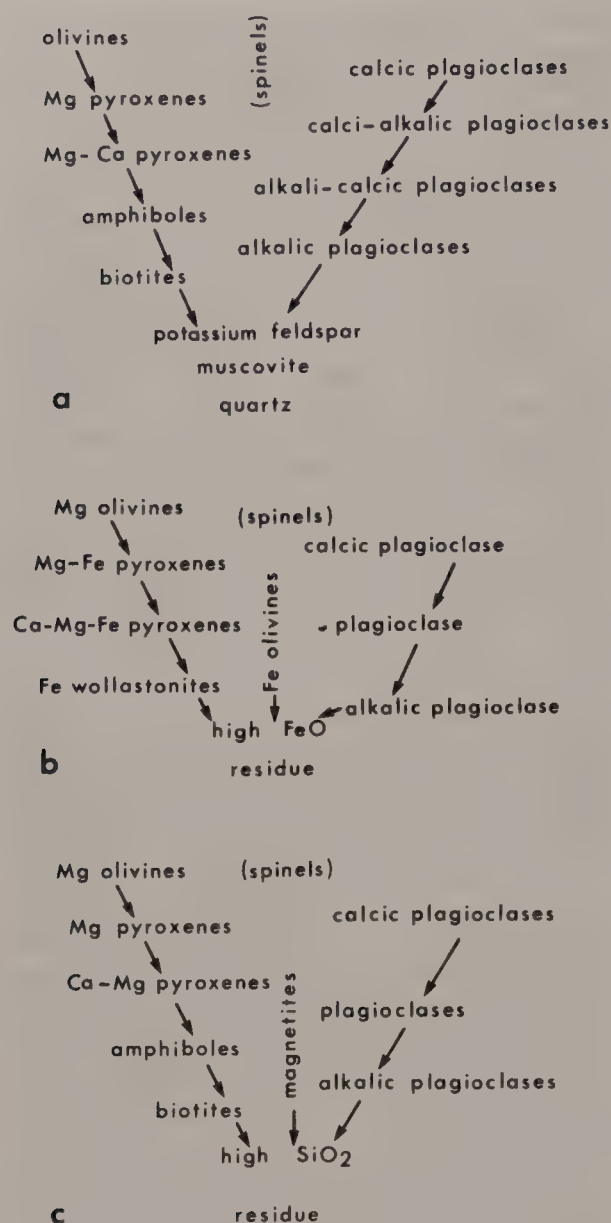


FIGURE 2. Reaction series for (a) subalkaline igneous rocks at low pressure (after Bowen, 1928), (b) in a closed system (after Osborn, 1979), and (c) under oxygen-buffered conditions (after Osborn, 1979.)

SiO₂, total Fe, Na₂O, and K₂O. Under oxygen-buffered conditions, magnetites would crystallize and the residual liquid that would result from fractional crystallization would be impoverished in MgO, CaO, and total Fe, and enriched in SiO₂, Na₂O, and K₂O (Fig. 2c). At medium (8–20 kb) and high pressure (>20 kb) most magmas have significantly different reaction relations and crystallization histories. For example, in basaltic liquids at medium pressures, olivine and plagioclase are usually not compatible phases, and pyroxene may coexist with spinels that precipitate early, whereas at high pressures garnet is usually the liquidus phase.

The first minerals to crystallize in a basaltic magma at low pressures are generally more dense than the adjacent liquid, and thus have a tendency to sink (gravitational differentiation). Stokes' law, which relates to the movement of small spheres in Newtonian fluids, demonstrates the importance of crystal size in determining the rate at which crystals move through a magma. Crystals suspended in non-Newtonian liquids (i.e., a Bingham plastic) must exceed a critical size before they can move through the magma, hence size is once again significant. When considering the movement of crystals in basaltic magmas, the plagioclase feldspars are of great interest, because at low pressures the liquidus plagioclases are more dense than most basaltic magmas, but if pressure is increased the density of the magma increases at a faster rate than that of the plagioclases. At a pressure of ca. 7 kb, plagioclase tends to float in normal basaltic magma. Even at low pressures, some minerals, such as leucite and the minerals of the sodalite group, may be less dense than the magma from which they precipitate. Wager and Deer (1939) suggested that as crystallization proceeded in the Skaergaard intrusion of E Greenland, the residual (rest) magma became more dense (i.e., more Fe-rich) than the precipitating plagioclase. Experimental studies have now confirmed this hypothesis.

Flowage magmatic differentiation

The composition of a magma may change while it is moving through the lithosphere. Some of the more important processes that produce these changes include (1) the crystallization and sinking of early-formed crystals, (2) the nucleation of crystals on the walls of the conduit, (3) wall-rock reaction, and (4) the escape of volatile components that are no longer soluble in a magma that has risen to higher levels and lower ambient pressures. The act of moving a liquid containing suspended crystals through a conduit tends to produce a concentration of crystals in the central part of the flow. This nongravitative method of concentrating crystals within a magma is called *flowage differentiation*.

Convection currents in magmas

In a convecting body of magma in which both the composition and temperature became relatively homogeneous, the most appropriate place for crystallization to occur would be at the base of the body of magma, where the pressure was highest (Jackson, 1961). In the case of the cooling of the Skaergaard intrusion, it has been suggested that there were two different sorts of

currents within the magma: one was a slow fairly continuous convective circulation, whereas the other was an intermittent density current of relatively small volume. Under these conditions, the crystals would nucleate in the cooler roof areas of a magma chamber, where they might either become an essential part of a density current or become entrained in the slower-moving convective currents, and eventually be deposited on the floor of the magma chamber. In order to account for the regular repetition of layers of rocks of different composition found in many layered intrusions, it has been proposed that the upward component in convection currents may provide a mechanism for separating minerals of different size and density. Jackson (1961) also introduced the concept of variable-depth convection. He postulated that most crystallization and settling takes place in a stagnant layer at the bottom of a magma chamber. Periodically, this layer heats up and becomes unstable, and the convective system operating above it increases in size and sweeps it aside. Fresh, cooler magma then forms a new stagnant layer, and starts a new cycle of crystallization and settling on the bottom of the magma chamber. It has also been proposed that many magma chambers develop a relatively thick mush of essentially unconsolidated crystals at their base, and the expulsion of interstitial, or intercumulus, liquid as the result of self-imposed filter pressing, promotes infiltration metasomatism and adcumulus growth in the pile of cumulus minerals (cf., Irvine, 1980).

Other factors

There are many more variables that may influence the course of magmatic differentiation in an essentially stationary body of magma, including (1) the size and shape of the magma chamber; (2) the position of the magma chamber within the lithosphere; (3) the shape as well as the sizes of the various solid phases that occur within a magma; (4) whether the solid phases aggregate into clusters; (5) changes in the density, polymerization and/or viscosity of the magma as crystallization proceeds; (6) variations in the rate at which different minerals nucleate; (7) whether, or not, the contents of the magma chamber are subjected to shock waves, movements, or pressures produced by tectonic processes; and (8) whether the materials in the magma chamber form an open, or closed, chemical system. With regard to whether a magma chamber is an open or closed chemical system, it is important to note that many magma chambers are of the periodically replenished, periodically tapped, continuously fractionating type. In addition, when a magma

becomes contaminated, the changes in chemical composition may produce sudden changes in the mineral, or minerals, being precipitated.

Other terms

Achistic, achistite—refers to a rock of a minor intrusion that has a composition equivalent to that of the parent magma, i.e., in which there has been no significant differentiation.

Accumulative phase—relates to that portion of a frequency distribution curve for a magmatic series that lies towards the low-SiO₂ side of the modal peak, presumably representing crystal accumulation as the result of differentiation.

Diachistic, diachistite—refers to those rocks of minor intrusions that appear to be, respectively, leucocratic and melanocratic differentiates of a common parent magma; also referred to as being *complementary*.

Differentiate—a rock formed by magmatic differentiation.

ERIC A. K. MIDDLEMOST

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Cross-references: *Cumulate textures*; *Double-diffusive convection*; *Geochemical modeling in petrogenesis*; *Hybridization*; *Intrusion—layered*; *Liquid immiscibility*; *Magma*; *Magmatic crystallization in open systems*; *Reaction principle and reaction series*; *Skaergaard intrusion*; *Textures of igneous rocks*.

MAGMATIC–TECTONIC CYCLE

The concept of a *magmatic–tectonic cycle* that has dominated the tectonic history of the Earth since Cambrian times was formulated by Stille (1940). The essentials of the concept are that every Phanerozoic fold mountain system went through four tectonic stages during orogeny, with each stage accompanied by a specific type of magmatism.

<i>Tectonic stage</i>	<i>Magmatic stage</i>
4. Cratonic	Final (basaltic volcanism)
3. Quasi-cratonic	Subsequent (acidic to basic volcanism and plutonism)
2. Orogenic	Synorogenic (granitic plutonism)
1. Geosynclinal	Initial (basic to intermediate volcanism)

D. R. BOWES

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MAGMATIC WATER—See Vol. IVA

MANTLE GEOCHEMISTRY—See Vol. IVA

MANTLE MINERALOGY—See Vol. IVB

MARBLE

Marble (Greek *marmaros*, sparkling) is a rock composed of minerals having essentially similar hardness and sufficiently dense texture to permit polishing. Scientific usage limits mineralogy to 95% carbonate (mainly calcite and dolomite) and the term is usually restricted to crystalline carbonate rock resulting from metamorphism. This conversion of limestone to marble by any metamorphic process is sometimes referred to as *marmorization* (marmarization, marmorosis, marmarososis) and was experimentally reproduced by Sir James Hall early in the 19th century by heating limestone under pressure in gun barrels.

Statuary marble is a uniformly glistening variety (e.g., Carrara, Italy). The presence of silicate minerals produces variegated marbles, including *ophicalcite*, in which green serpentinized forsterite or diopside is set in a white carbonate matrix. Some of these varieties are much prized for ornamental purposes and much used in architecture. Commercial usage of the term marble also covers ornamental varieties resulting from the precipitation of calcite and aragonite (as travertine), together with chalcedony (SiO_2), by hydrothermal waters or hot springs, e.g., oriental alabaster (Mexican onyx that originally faced the Pyramids of Egypt) is travertine with bands of chalcedony. Alabaster, a fine-grained compact variety of gypsum, is also known commercially as marble.

Marble occurs in most regionally metamorphosed terranes from greenschist to granulite facies and is found in contact (thermal) aureoles around igneous intrusions. There is an age control in distribution, with calcite marbles being more common in rocks younger than ca. 1.0 Ga, whereas dolomite marbles predominate in older rocks. A major exception to this is the white calcite marble of the Vindhyan Hills of central India that was used in the construction of the Taj Mahal and is early Proterozoic–late Archaean in age.

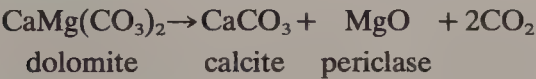
Mineralogy

Pure calcite marble simply undergoes recrystallization during progressive metamor-

TABLE 1. Mineralogy of marbles

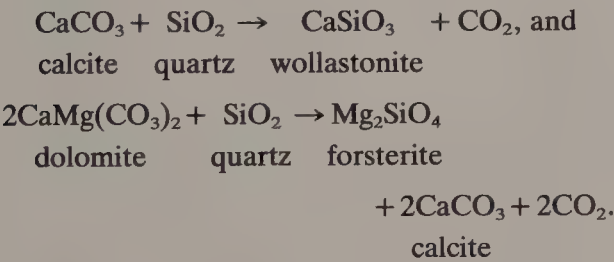
Impurity	Calcite marble	Dolomite marble
—	Calcite	Dolomite, calcite, brucite, periclase
SiO ₂	Wollastonite, spurrite	Diopside, forsterite, humite group, tremolite, talc, serpentine, anthophyllite, monticellite, merwinite
FeO, SiO ₂ Al ₂ O ₃ , SiO ₂	Magnetite Grossular, melilite group, plagioclase, scapolite, epidote group	Actinolite, magnetite Grossular, spinel, melilite group
FeO, SiO ₂ , Al ₂ O ₃	Andradite	Spinel, andradite, almandine, hornblende, pargasite

phism, even in the highest grades (facies) with the development of granoblastic texture. In the case of dolomite, decarbonation accompanies metamorphism and when dolomite is thermally metamorphosed,



with the periclase generally being hydrated to brucite (Mg(OH)₂).

The presence of other impurities, particularly quartz, permits a more varied mineral assemblage to develop by reactions such as



Progressive stages of metamorphism in aureoles of contact (thermal) metamorphism are marked by the appearance or disappearance of some mineral or assemblage (Bowen, 1940).

In both thermal (contact) and dynamothermal (regional) metamorphism, the presence of particular phases is determined by the nature of the impurities (Table 1), the pressure and temperature, and the activity of SiO₂, H₂O, and CO₂. A very diverse assemblage of *calc-silicate rocks* (sometimes referred to as *skarns*) is seen in nature, varying from impure marbles to rocks in which all the carbonate has been used up in metamorphic reactions (Harker, 1952; Winkler, 1967; Turner, 1981).

Other terms

Calciiphyre—a marble containing conspicuous crystals of Ca- and/or Mg-silicates such as forsterite, pyroxene and garnet.

Cipolin—a term used in France for any crystalline limestone.

Cipolino—a siliceous marble containing micaeous layers.

Pencatite—a marble containing periclase or brucite and calcite in approximately equal molecular proportions formed by contact (thermal) metamorphism of magnesian limestone.

Predazzite—a marble similar to pencatite but in which the proportion of brucite is less than that of calcite.

D. R. BOWES

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Cross-references: *Aureole*; *Contact (thermal) metamorphism*; *Facies of thermal metamorphism*; *Metamorphic reactions*; *Metamorphism*; *Regional metamorphism*; *Skarn*.

MEDITERRANEAN SUITE

The *Mediterranean suite* is a major group of igneous rocks, characterized by high alkali contents with K dominant over Na and typified by the volcanic rocks of south-central Italy, including those of Vesuvius. Mineralogically leucite, sanidine, biotite, and titanaugite or diopside with aegirine are the most common constituents in an assemblage of tephritic phonolites, phonolitic tephrites, and tephritic

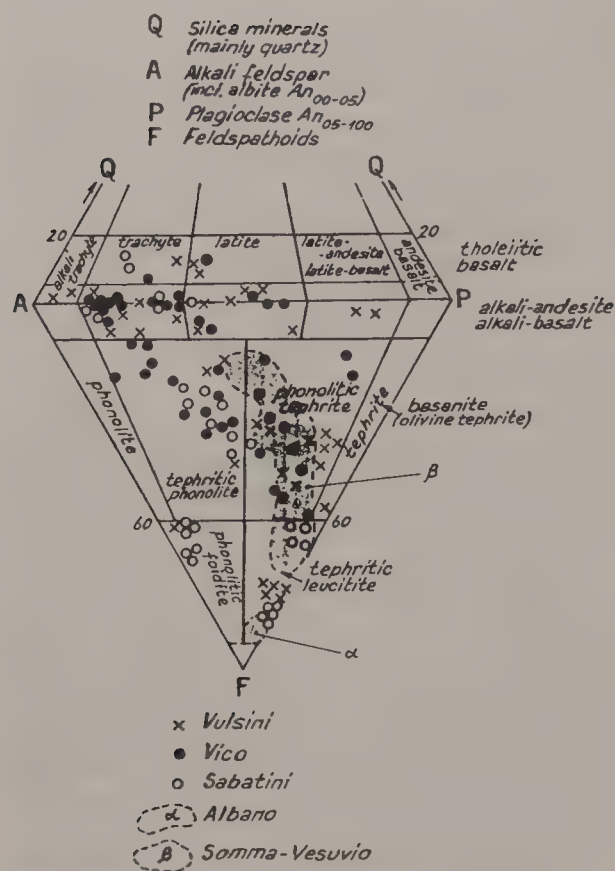


FIGURE 1. O-A-P-F plot of compositions of volcanic rocks from five districts in south-central Italy from Vulsini southeastwards (ca. 300 km) to Somma-Vesuvius.

leucitites (Fig. 1). These are associated with olivine-bearing trachybasalts, trachyandesites, and trachytes and a restricted number of plutonic masses of malignite, shonkinite, and borolanite. The concepts of consanguinity (i.e., showing characteristics that are consistently the same or show gradual change) and petrographic province, have been applied in relation to this suite in a manner comparable to that in the case of the Pacific and Atlantic suites. However, while no single geological-geotectonic environment has been proposed for the Mediterranean suite, mantle-derived metasomatic fluids are recognized as playing an essential role.

Rock associations characterized by high K in other parts of the world that have been included as representatives of the Mediterranean suite vary in age from Palaeozoic to Recent. Included amongst these are those of Leucite Hills, Wyoming, Highwood Mts., Montana, U.S.A., Birunga and Toro Ankole, Uganda and W Kimberley, W Australia.

E. LOCARDI

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Cross-references: Alkaline rocks—undersaturated; Feldspathoidal rocks—genesis; Leucitite; Melilitite; Pacific and Atlantic suites; Petrographic province; Trachyte.

MELILITITE

Melilitite is an undersaturated, feldspar-free, extrusive or hypabyssal intrusive rock composed essentially of melilite and clinopyroxene (usually titaniferous augite). The name was first used by A. Lacroix in 1896 (Holmes, 1928), but Johannsen (1938) ascribes its first use to G. T. Prior in 1902 in describing a basaltic rock from the Nyando River, E Africa, related to nepheline but having the nepheline almost wholly replaced by melilite. Feldspathoids may be present, yielding nepheline melilitite and leucite melilitite, respectively. Where the feldspathoids become dominant, these rocks are referred to as melilite nepheline and melilite leucite, respectively.

Where olivine is present in more than accessory proportions, the melilite-pyroxene-olivine volcanic rocks have formerly been called melilite basalts, but would be better described as olivine melilitites and the feldspathoid-rich varieties olivine-melilite nephelinites (and leucitites). Common accessories in melilitites are opaque ores, sphene, perovskite, apatite, and biotite.

There has been much confusion in the use of the terms *melilitite* (essentially melilite and clinopyroxene), *melilithite* (proposed originally by F. Y. Loewinson-Lessing in 1901 for rocks composed exclusively of melilite, for which the name melilitite would be more appropriate), and *melilitholith*, a name proposed by Lacroix in 1933 (see Tröger, 1935) for the group of rocks consisting almost entirely of melilite and the mafic minerals (augite, olivine, and biotite). The melilitholiths are *uncompahgrite* (melilite, augite), *alnöite* (melilite, biotite, augite, olivine), *melilitite* (melilite, augite), olivine melilit-

ite (melilite, augite, olivine), and *coppaelite* (melilite, diopside, phlogopite). Johannsen (1938) uses melilitite for a rock composed of melilite and augite without olivine, so of the melilitholiths he would regard uncompahgrite and melilitite as melilitites, olivine melilitite as melilite basalt; alnöite and coppaelite, which contain essential biotite, can be regarded as biotite melilitites. Of these, the following rocks are most closely related to melilitite.

Uncompahgrite is a coarse-grained equivalent of melilitite. It was named by E. S. Larsen and J. F. Hunter in 1914 from the Iron Hill Ring Complex, Colorado, U.S.A. Coarsely crystalline and xenomorphic to hypidiomorphic granular, it is somewhat porphyritic, consisting of approximately 70% melilite, together with pyroxene, magnetite, sometimes biotite, and accessory apatite, calcite, perovskite, anatase, and melanite. Other coarse-grained melilite-rich rocks are *turjaite* (first described by W. Ramsay in 1921 from the Kola Peninsula, U.S.S.R.), and *okaite* (first described by J. Stansfield in 1923 from near Quebec), both of which are of similar mineralogical composition. Turjaite (Hatch et al., 1972) has large conspicuous crystals of titaniferous biotite in a melilite matrix, which constitutes half the rock. Nepheline is an essential component; apatite, magnetite, and perovskite are accessories and some specimens contain olivine, melanite, and calcite. In okaite the place of nepheline is taken by haüyne.

Alnöite, a porphyritic dyke rock first described by H. Rosenbusch in 1887 from Alnö Island, Sweden. It contains phenocrysts of biotite, olivine and augite in a groundmass of melilite and augite with perovskite and other accessories (Meyer and Villar, 1984). When nepheline is an essential component the rock is termed a *bizardite*.

Hatch et al. (1972) have drawn attention to the difficulties of nomenclature of the feldspar-free lavas, including melilitite, and suggest that the name *melmafite* (*melilitite*) be used for the rocks consisting of melilite plus mafic minerals, which are currently called melilitite and melilite basalt (= olivine melilitite).

The classification and nomenclature of the melilitic rocks is set out by Streckeisen (1980) and the petrogenesis of the melilitites is dealt with by Frey et al. (1978), Onuma and Yamamoto (1976), and Ibarrola and Martorell (1973).

The close association of olivine melilitites, kimberlites, and carbonatites in Cape Province, S Africa has led to the suggestion of a genetic relationship (e.g., McIver and Ferguson, 1979). However, significant differences are demonstrated by Boctor and Yoder (1986): kimberlites

are generally richer in MgO and poorer in CaO and Na₂O than olivine melilitites and there are differences in the chemistry of phenocryst minerals and in the groundmass mineralogy. However, both are products of melting deep in the mantle of carbonated peridotite and similarities exist between the residual liquids that evolve in the final stages of crystallization of olivine melilitite and kimberlite—the liquids are apparently enriched in Ca, REE, Ti, and other incompatible elements and in CO₂ and H₂O. It is likely that the ratio of volatiles and their degree of retention are the principal variables in the relationship of the magmas that eventually yield the low-*P* assemblages characteristic of olivine melilitite and kimberlite. Other variants may result from fractionation or from varying degrees of partial melting of a carbonated mantle peridotite.

R. T. PRIDER

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Cross-references: *Alkaline rocks—undersaturated; Basalt; Feldspathoidal rocks—genesis; Kimberlite; Lamprophyre; Leucitite; Nephelinite; Peralkaline igneous rocks; Rift valley volcanism; Upper mantle—experimental petrology.*

METAMORPHIC CONVERGENCE

Metamorphic convergence is the process by which similar-looking metamorphic rocks or rock complexes are formed from different parent materials.

As a result of metamorphism, the same parent rock may develop into rocks of strikingly different appearance by recrystallization under different metamorphic conditions; e.g., basalt may be metamorphosed to greenschist, amphibolite, amphibole–plagioclase(–pyroxene) granofels, pyriclasite, amphibole–plagioclase cataclastite, etc. This phenomenon might be called metamorphic divergence, but such an expression is not in common use. Alternatively, quite different parent materials may, by metamorphic processes, be transformed into similar rocks. Parent rocks that are different in structure but similar in chemical bulk composition, are more likely to “converge” into similar metamorphic products than rocks that have large initial differences in chemical composition.

Amphibolites. Foliated amphibole–plagioclase rocks (amphibolites) may form from gabbros and from basic lavas and pyroclastic products. The latter may be deposited or redeposited in water and thereby receive some carbonate, pelitic, or quartzofeldspathic admixture. Local geologic correlations and the presence of relict structures may be used to discriminate between the different types of basic parent rocks. It has also been proposed that marls (mixtures of pelitic material and carbonates) could be suitable parent material for (para-) amphibolites. However, major- and trace-element data from suites of amphibolites generally reveal chemical variation patterns different from those expected for marls, but conforming to those for basic magmatic rocks (Fig. 1; Leake, 1964; van de Kamp, 1968). This indicates that metamorphic convergence was not a major factor in amphibolite formation, at least in many regions.

Gneisses. Foliated quartzofeldspathic rocks

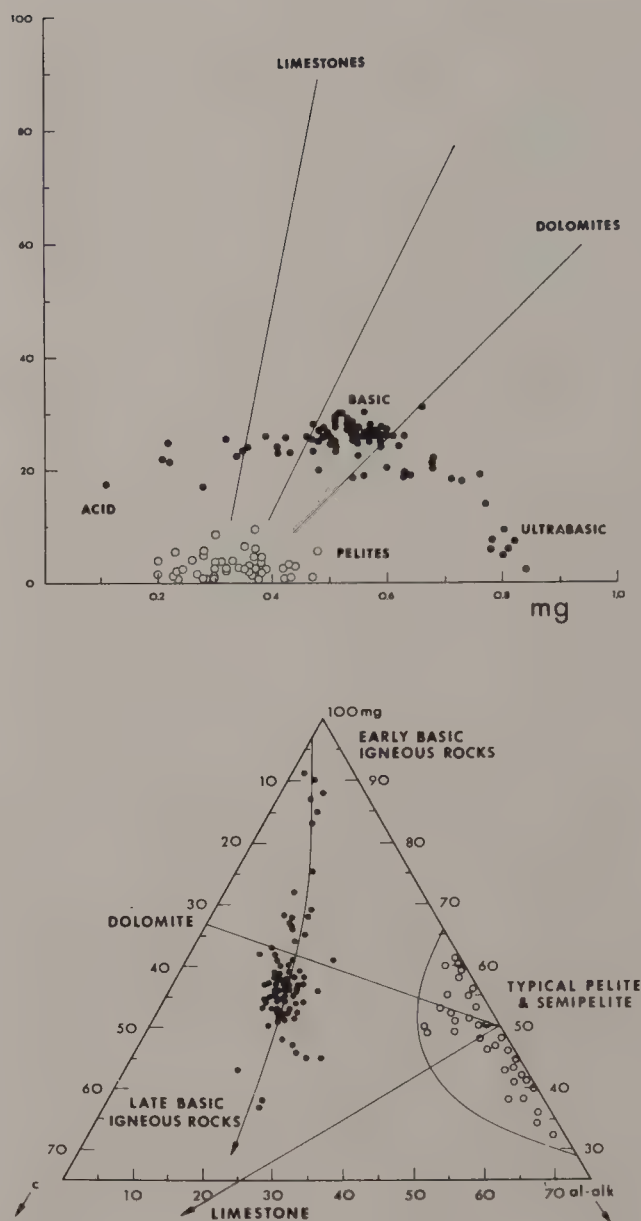


FIGURE 1. Plots of Niggli values used to discriminate between a suite of metamorphosed igneous rocks (●) and standard carbonate and pelitic rock compositions (○). (After van de Kamp, 1968.)

may be formed from arkosic sediments, acid volcanic rocks, or granitic rocks. Field relations and relict structures may provide indications for one of these protoliths. However, care should be taken in interpreting a layered structure as proof of supracrustal origin. Isoclinal folding with concomitant recrystallization may produce layered gneisses, and also shearing and cataclasis accompanied or followed by recrystallization may, and often does, produce layering (Myers, 1978). Many banded gneisses, after being labeled as metasediments for years, have been shown to be of igneous parentage (e.g., Zeck and Mallin, 1974).

Chemical characteristics have been used to

distinguish sediment-derived paragneisses from igneous-derived orthogneisses with the occurrence of Al-silicate and corundum in the norm often taken as an indication for a sedimentary origin. Shaw (1972) compared the chemical composition of large numbers of igneous and sedimentary rocks and was able to put forward a differentiation factor (DF) to distinguish sedimentary from igneous derived gneisses ($DF = 10.44 - 0.21SiO_2 - 0.32Fe_2O_3$ (total Fe) $- 0.98MgO + 0.55CaO + 1.46Na_2O + 0.54K_2O$). A negative value for this parameter would indicate a sedimentary parentage and a positive value an igneous parentage.

Marbles. Practically all marbles (metamorphic carbonate rocks) are metasediments. Those representing metacarbonatites are very rare but can be recognized on the basis of very high proportions of Ba, Sr, Y, Nb, Ta, and REE such as known in carbonatites.

Eclogite. Two types of eclogitic rock of quite diverging origin but similar appearance occur within a Precambrian migmatite complex in W Norway (Griffin and Raaheim, 1973). One type, which occurs as lenses and layers within migmatites and is often associated with metasedimentary rocks such as marbles and quartzites, is interpreted as having formed syngenetically with the host essentially from basaltic volcanic rocks. The other type was formed within gabbroic sills that had intruded the gneiss complex after the formation of the first type. Many of these intrusive bodies have developed a 20–200 cm wide marginal zone that consists largely of garnet and jadeite-rich clinopyroxene. Farther from the margin are rocks with coronas in which orthopyroxene, clinopyroxene, and garnet replace olivine and plagioclase, while in the central parts of the bodies magmatic, subophitic textures may be preserved. These basic bodies are interpreted as representing magma that was emplaced as sills at a pressure of 4.5–9 kb and, after recrystallization and cooling to the temperature of the country rocks (700°C), the eclogite mineralogy was produced. At kinetically favorable places the transformation has nearly gone to completion but at others the various reaction stages are preserved in coronas. Both types of eclogitic rocks share a common retrogressive imprint probably imposed during subsequent regional uplift (Fig. 2).

Prograde vs. retrograde products. In places where progressive structural and mineralogical changes cannot be seen, it is often difficult to discriminate between amphibolite facies complexes of prograde metamorphic origin and those that represent retrogressive, originally granulite facies complexes such as those of the Lofoten–Vesteraalen Islands, N Norway

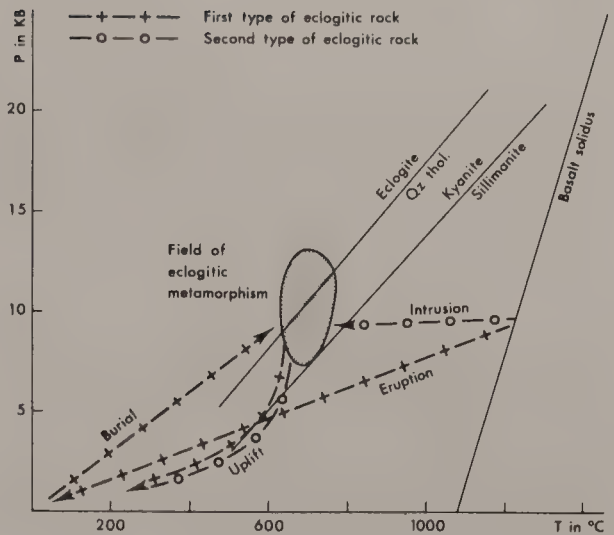


FIGURE 2. Metamorphic convergence illustrated by the P - T paths of two types of eclogitic rocks in a migmatite complex, W Norway. (After Griffin and Raaheim, 1973.)

(Griffin and Heier, 1969) and Godthaab–Fiskenaesset–Frederikshaab area, W Greenland (Kalsbeek, 1974). Use of geochemical parameters might solve the problem: e.g., medium–high pressure granulite facies rocks should be characterized by low contents of incompatible elements like K, Rb, Cs, U, and Th.

H. P. ZECK

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Cross-references: *Amphibolite; Carbonatite; Eclogite; Gneiss; Granulite; Magmatic crystallization in metamorphic environments; Marble; Metamorphic*

facies; Petrochemical calculations; Regional metamorphism.

METAMORPHIC ENVIRONMENTS—
CHEMICAL MOBILITY—See Vol. IVa

METAMORPHIC FACIES

A *metamorphic facies* is a group of rocks which for identical bulk composition exhibits an identical mineral composition, but whose mineral composition for varying bulk composition varies according to definite laws. Rocks that belong to the same metamorphic facies and reached equilibrium under the same set of physical conditions are referred to as being *isofacial*. This concept was developed by P. Eskola from the early work of F. Becke and V. M. Goldschmidt which had showed that rocks varied in mineral assemblage as the metamorphic environment changed. Each facies that he defined was thus a series of mineral assemblages covering the known range of normal metamorphic rock chemistry, some compositions having parageneses common to several facies. Rock texture and field disposition (i.e., regional vs. local extent) were not considered and play no part in determining metamorphic facies. Although temperature and pressure of formation were not implied, it was possible to make provisional correlations with temperature and pressure for the nine facies he described (Fig.1).

The increase in knowledge of metamorphic parageneses and particularly the vast increase in knowledge of the physical conditions of formation of mineral suites by experimental studies has enabled this classification to be widened in scope to include several more facies and for the original boundaries to be defined accurately in relation to mineral stability fields. Turner (1958, 1968, 1981) synthesizes this reappraisal and (1958, p. 18) defines a metamorphic facies as a series of mineral assemblages conforming to the following specifications.

- 1. Some or all the assemblages are commonly associated in space and in time (e.g., in a contact aureole or in a mappable metamorphic zone). The series of assemblages tend to recur in different regions and in rocks of different age.
- 2. The mineralogical composition of each assemblage is strictly a function of the bulk chemical composition of the rock as it now exists. Gradation from one assemblage into another throughout the facies can be correlated with serial variation in chemical composition (e.g., correlation between mineralogy and chemical composition as illustrated by the ACF diagram).
- 3. The total number of mineral phases occurring as essential constituents of common rocks in the facies is relatively small—perhaps 12. Each assemblage consists of even fewer essential minerals—generally between 2 and 6.
- 4. The minerals of any particular assemblage show no textural or other evidence of actual replacement.

Two tentative assumptions are made in order to interpret genetically the facies so defined: (1) rocks belonging to any particular facies were formed within the same range of physical conditions; (2) the constituent mineral assemblages were stable at the time they were formed.

As at present defined, metamorphic facies imply the physical fields necessary to produce the “type” mineral facies, but are not defined on the basis of temperature and pressure estimates. A wide range of other mineral parageneses lies within any one facies that may not include the named minerals. The term “critical mineral” has been used to refer to a particular mineral, or one of a mineral association, that is stable only under the conditions of one metamorphic facies.

The subdivision and naming of the various metamorphic facies is now in a state of some confusion (cf., Fyfe and Turner, 1966, see Hyndman, 1972; Hietanen, 1967; Winkler,

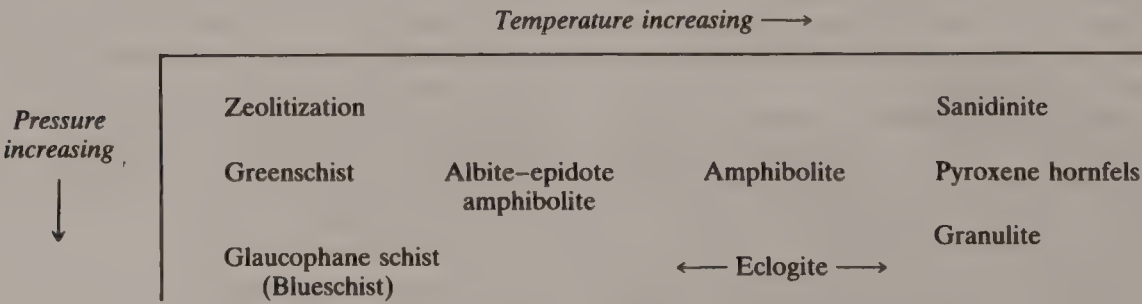


FIGURE 1. Facies of P. Eskola in relation to temperature and pressure.

1967; Lambert, 1972; Turner, 1981) with statements by the various authors differing in some respects from the others. In the 1950s and early 1960s there was a proliferation of new facies and subfacies suggested by various authors, which became so cumbersome that there has been a reaction to the rigorous definition of subfacies and even proposals to discard the facies concept entirely (Winkler, 1970). In his latest reassessment, Turner (1981) suggests four contact-metamorphic and seven regional-metamorphic facies with transitional facies between, since "to invent new names to cover these transitional facies... would result in falsely presenting a picture of sharp interfaces boundaries where none exist."

Turner (1958) suggested that the epidote amphibolite facies of Eskola should be regarded as a subfacies of the greenschist facies, principally because albite up to An_7 is stable throughout the greenschist and epidote amphibolite facies but at the amphibolite facies boundary the calcic-aluminum silicates become unstable and more basic plagioclases start to appear. Chlorite is also still stable in pelitic and basic rocks and thus the parageneses are very similar to the greenschist facies. This suggestion has been adopted in many textbooks, yet many authors (e.g., Hietanen, 1967; Lambert, 1972;

and the Japanese school) still regard it as one of the major facies (Fig. 2).

Early compilations regarded low-grade contact (thermal) metamorphic assemblages as subfacies of the greenschist and amphibolite facies, but the division of metamorphic facies into two major groups, those of contact (thermal) metamorphism ("hornfels facies") and those of regional metamorphism, has met with almost universal acceptance.

The most complete systems of metamorphic facies classification are those of Turner (1958) and Winkler (1967), but some of the deficiencies of these have subsequently been recognized by those authors (Turner, 1968, 1981; Winkler, 1970), while Lambert (1972) draws attention to the impossibility of drawing accurate facies boundaries based on several different assemblages. Instead, he suggests that major divisions are defined in relation to basaltic rocks but that the subdivision of the amphibolite facies (which occupies a large and critical area of the petrogenetic grid) be defined in relation to alumina-saturated pelites. Although the major facies, as generally accepted, are defined on the basis of basaltic compositions, it is possible to erect mineral facies on any consistent assemblage; this has been done by Thompson and Thompson (1976) for pelitic

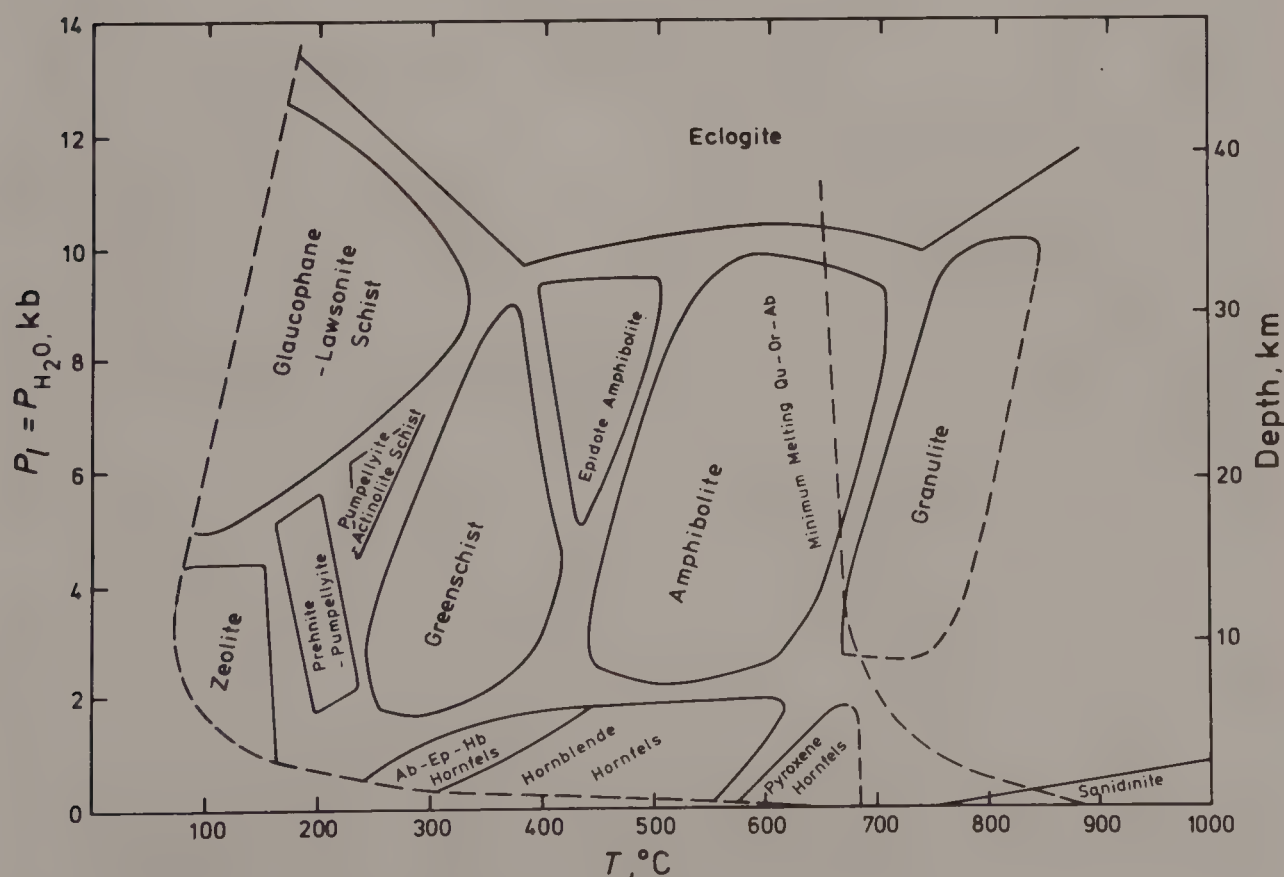


FIGURE 2. Metamorphic facies. (Adapted from Turner, 1968.)

assemblages and they recognize 36 mineral facies of quartz-muscovite- and/or K-feldspar-bearing pelitic rocks, which divides up the petrogenetic grid at a rather uniform spacing.

A complete reappraisal and redefinition of all the metamorphic facies of Eskola and Turner has been made by Sobolev and his co-workers (1972–1975). Their scheme attempts to define the limiting mineralogical assemblages for each of their facies and subfacies, on the basis of pelitic assemblages over much of the grid. Because the facies limits vary in many cases from the Eskola facies, the Russian authors have suggested new names for all the major facies (Fig. 3). The names proposed by these

authors differ in the different volumes: generally in the following list the more distinctive names have been used. For example, it seems wrong to use amphibolite facies for the biotite-sillimanite gneiss facies as they do in volume 3 (1973), as it represented only one subfacies (of six) of the amphibolite facies as defined by other authors.

The majority of petrologists, however, now simply use a rather generalized facies scheme (such as Fig. 2) to give a general idea of facies involved, but as it is becoming increasingly possible, by detailed phase chemistry, to estimate the temperature and pressure at which a particular rock recrystallized, there is no need to have semantic arguments about the names of

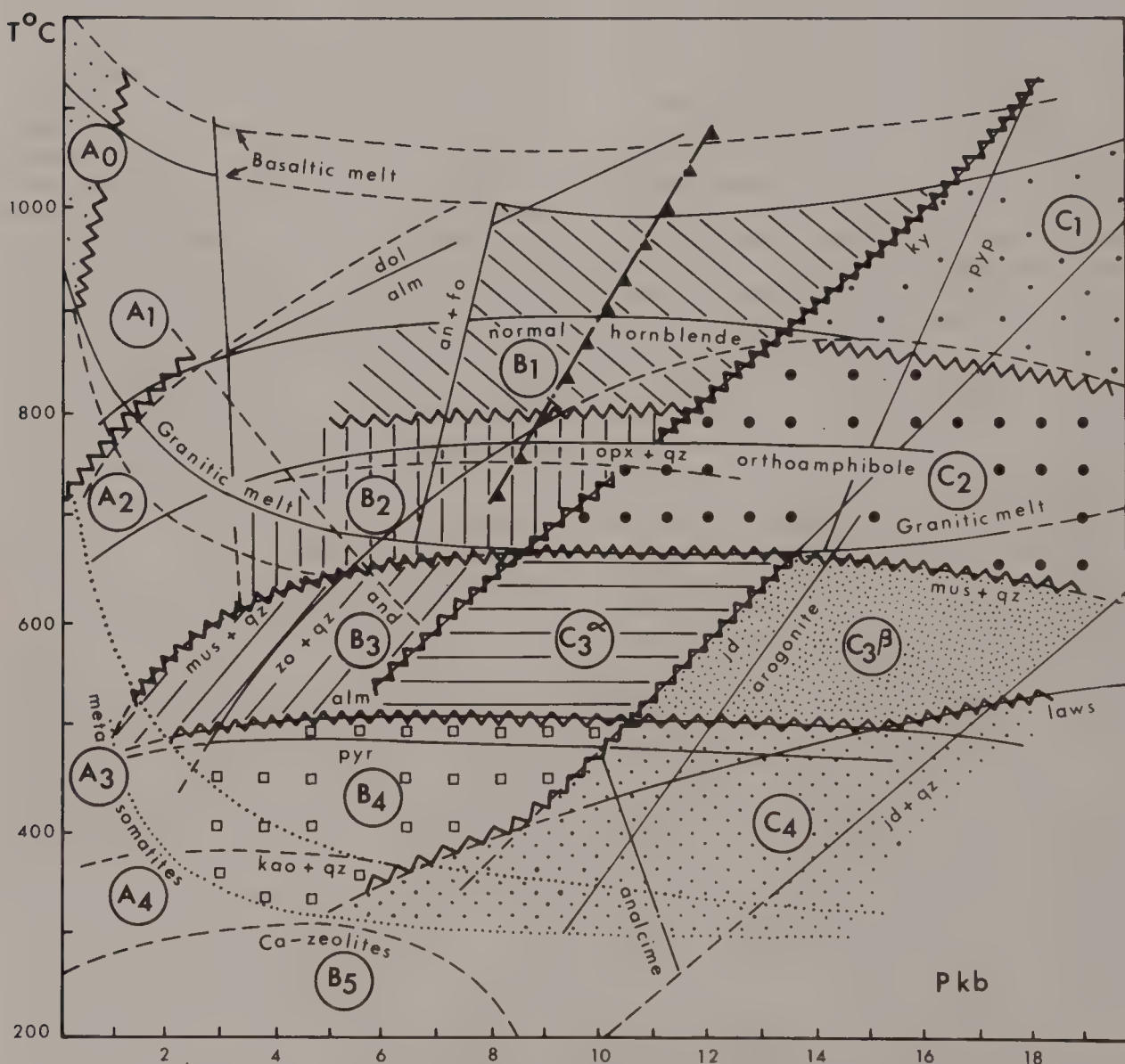


FIGURE 3. Facies scheme of Sobolev (1972). A₀, spurrite-merwinite; A₁, pyroxene hornfels; A₂, amphibole hornfels; A₃, muscovite hornfels; A₄, hydrothermal and metasomatic rocks associated with contact aureoles; B₁, two-pyroxene gneiss (granulite); B₂, biotite-sillimanite gneiss (amphibolite); B₃, muscovite-staurolite schist (epidote amphibolite); B₄, greenschist; B₅, zeolite; C₁, eclogite; C₂, kyanite gneiss; C₃, kyanite-muscovite schist (glaucophane-almandine); C₄, jadeite-lawsonite-glaucophane (glaucophane schist.)

facies or their precise limits when the limits of stability of a particular rock have been determined.

The following list of metamorphic facies attempts to review facies and subfacies that have been proposed, although many are of

doubtful validity. The facies suggested by Sobolev and his collaborators, although not exactly comparable to those of Eskola and Turner, are given next to the facies they most closely approximate.

Facies of contact metamorphism

Load pressure is low (0.1–3 kb) and water pressure is highly variable: the facies are listed in order of increasing temperature.

Facies of Turner (1968) and other authors

- (1) Albite–epidote hornfels facies
 - (a) Albite–epidote (Miyashiro, 1973)
 - (b) Actinolite–calcic plagioclase (Miyashiro, 1973)
- (2) Hornblende hornfels facies
 - (a) Muscovite
 - (b) K-feldspar
 } (Winkler, 1967)
- (3) Pyroxene hornfels facies
(= orthopyroxene subfacies of K-feldspar hornfels facies; Winkler, 1967)
- (4) Sanidinite facies
 - (a) Monticellite–melilite
 - (b) Larnite–merwinite–spurrite

Facies of Sobolev (1972–1975)

- Muscovite hornfels facies
- Amphibole hornfels facies
- (a) Andalusite
 - (b) Sillimanite
- Pyroxene hornfels facies
- (a) Grossular
 - (b) Monticellite–melilite (= wollastonite–gehlenite–anorthite)
- Spurrite–merwinite facies
- (a) Monticellite–spurrite–tilleyite
 - (b) Merwinite–calcite

Facies of regional metamorphism

Low-temperature facies (burial metamorphism of some authors). The temperature is low (<350°C) and water pressure increases with or even exceeds load pressure which is low to very high: the facies are listed in order of increasing pressure.

Facies of Turner (1968) and other authors

- (5) Brownstone facies (Cann, 1979)
- (6) Zeolite facies (= laumontite–prehnite–quartz facies; Winkler, 1967)
 - (a) Heulandite (Coombs, 1960)
 - (b) Laumontite (Coombs, 1960)
 - (c) Muscovite–chlorite–dolomite–quartz (McNamara, 1965, see Hyndman, 1972)—higher P_{CO_2} than (a) and (b)
- (7) Prehnite–pumpellyite facies (Hashimoto, 1966, see Hyndman, 1972)
 - (a) Quartz–prehnite–pumpellyite
 - (b) Muscovite–chlorite–epidote–calcite (McNamara, 1965, see Hyndman, 1972)—higher partial pressure of CO_2 than (a)
- (8) Pumpellyite–actinolite schist facies (Hashimoto, 1966, see Hyndman, 1972)
- (9) Glaucophane–lawsonite schist facies (= lawsonite blueschist facies)
 - (a) Lawsonite–albite (McKee, 1962, see Hyndman, 1972)
 - (b) Lawsonite–jadeite–quartz (McKee, 1962, see Hyndman, 1972)
 - (c) Almandine–lawsonite–glaucophane (De Roever, 1950, see Turner, 1981)

Facies of Sobolev (1972–1975)

“Zeolite facies” (not recognized as a true facies of regional metamorphism)

Greenschist facies

- (a) Pumpellyite–stilpnomelane

Glaucophane schist facies

- (a) Lawsonite–albite
- (b1) Crossite–actinolite
- (b2) Glaucophane–epidote–chlorite
- (c) Pumpellyite–lawsonite–glaucophane

Facies of regional metamorphism (*continued*)Facies of Turner (1968)
and other authors

- (d) Epidote–glaucophane–jadeite–quartz
(De Roever, 1950, see Turner, 1981)

Facies of Sobolev (1972–1975)

- (d) Epidote–lawsonite–glaucophane
- (e) Quartz–jadeite–glaucophane
- (f) Almandine–lawsonite–glaucophane

Medium- to high-temperature facies. Temperature 400–700°C and pressure is low to high; water pressure is variable but generally increases with load pressure: the facies are listed in order of increasing temperature and pressure.

Facies of Turner (1968)
and other authors

Facies of Sobolev (1972–1975)

(10) Greenschist facies

Greenschist facies

- (a) Quartz–albite–muscovite–chlorite–epidote
- (b) Muscovite–chlorite–epidote–calcite
(McNamara, 1965, see Hyndman, 1972)—
higher P_{CO_2} than (a)
- (c) Quartz–albite–epidote–biotite
- (d) Glaucophanitic greenschist “facies”
(Winkler, 1967) probably = glaucophane–
epidote (Miyashiro and Seki, 1958, see
Turner, 1981)

- (b) Epidote–muscovite–chlorite

- (c) Biotite–chlorite

(11) Epidote amphibolite facies

Muscovite–staurolite schist—also called andalusite–
(sillimanite)–muscovite

- (a) Quartz–albite–epidote–almandine
- (b) Quartz–andalusite–plagioclase–chlorite—
lower P than (a)

(12) Amphibolite facies—divided into two “facies”
by Winkler (1967)

Cordierite amphibolite

- (a) Andalusite–cordierite–muscovite
- (b) Sillimanite–cordierite–muscovite–
almandine
- (c) Sillimanite–cordierite–orthoclase–
almandine

Biotite–sillimanite gneiss facies

Almandine amphibolite

- (d) Staurolite–almandine
- (e) Kyanite–almandine–muscovite
- (f) Sillimanite–almandine–
orthoclase

Kyanite–muscovite schist facies
Kyanite gneiss facies

Subfacies (d)–(f) form at higher P than those of (a)–(c). Lambert (1972) shows that there are 12 assemblages possible in aluminous pelites within the amphibolite facies based on the stability of andalusite–kyanite–sillimanite, chloritoid–staurolite, garnet, and muscovite–orthoclase.

Extremely high-temperature and high-pressure facies. One of the two variables, temperature and pressure, is very high and the other moderate to very high. The presence of water is important in the genesis of these rocks, since rocks that are originally anhydrous may recrystallize in these facies at lower temperatures and pressures than are otherwise required.

Facies of Turner (1968)
and other authors

Facies of Sobolev (1972–1975)

(13) Granulite facies

Two-pyroxene gneiss facies

- (a) Biotite–cordierite–almandine (de Waard, 1965)
- (b) Hornblende–orthopyroxene–plagioclase
(de Waard, 1965)

Facies of regional metamorphism (*continued*)

*Facies of Turner (1968)
and other authors*

- (c) Hornblende–clinopyroxene–almandine (de Waard, 1965)
- (d) Cordierite–almandine (de Waard, 1965)
- (e) Orthopyroxene–plagioclase (de Waard, 1965)
- (f) Clinopyroxene–almandine (de Waard, 1965)

Subfacies or facies of other authors may be correlated with these.

Hornblende granulite a, b, c
Pyroxene granulite d, e, f
Cordierite granulite a, d
Garnet granulite c, f

(14) Eclogite facies

- (a) Group A: pyrope content of garnet >55% (eclogites in kimberlites and peridotites—Coleman et al., 1965)
- (b) Group B: pyrope 30–55% (eclogites in gneisses—Coleman et al., 1965)
- (c) Group C: pyrope <30% (eclogite in glaucophane–lawsonite schists—Coleman et al., 1965)

Facies of Sobolev (1972–1975)

Eclogite facies

Notes

1. Miyashiro (1973) divides a subfacies of the albite–epidote hornfels facies that is characterized by the more calcic plagioclase (owing to loss of zoisite). The other rocks in this facies he states are of greenschist mineralogy.

2(a–b). Winkler (1967) replaced the pyroxene hornfels facies with the K-feldspar–cordierite hornfels facies, dividing this into two by the incoming of orthopyroxene. The lower grade subfacies is thus included by other authors in the hornblende hornfels facies. The distinction is thus made within the hornblende hornfels facies on the presence of muscovite or K-feldspar.

6(c), 7(a), 10(b). These subfacies have been proposed by McNamara (1965) as equivalents to the normal zeolite–metagraywacke–greenschist facies transition where there is a high relative concentration of CO₂ in basic and pelitic rocks, although other authors regard these parageneses as indicative of greenschist facies. Turner (1968) suggests that such parageneses may be found side by side and such small variations should not be used to erect subfacies.

7, 8. These two facies are differentiated here following Hashimoto (1966, see Hyndman, 1972). Many authors regard them as subfacies. Schermerhorn (1975) suggests that the two be

combined in one pumpellyite facies. Many authors use prehnite–pumpellyite as the name of the facies (Turner, 1981).

9, 10(d). Subfacies of the glaucophane–lawsonite schist facies seem to vary widely from area to area and the facies grades up into greenschists usually with glaucophane. These were included in the glaucophane schist facies before the use of lawsonite as the critical mineral of the facies. The glaucophanitic greenschist “facies” (10d) of Winkler (1967) is of this type, although critical mineral stability relationships to distinguish it from greenschists are not recorded.

12(a), 13(a),(d). Dallmeyer (1972) has shown that near the amphibolite facies–granulite facies boundary the assemblages containing cordierite are very strongly affected by rock chemistry and that cordierite is an unsuitable mineral for defining subfacies.

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Cross-references: *Amphibolite facies*; *Brownstone facies*; *Burial metamorphism*; *Contact (thermal) metamorphism*; *Eclogite facies*; *Epidote amphibolite facies*; *Facies of thermal metamorphism*; *Glaucophane schist facies*; *Granulite facies*; *Greenschist facies*; *Lawsonite-albite-chlorite facies*; *Prehnite-pumpellyite facies*; *Pumpellyite-actinolite schist facies*; *Regional metamorphism*; *Zeolite facies*.

METAMORPHIC FACIES SERIES

Within metamorphic terranes, a sequence of zones of gradually increasing grade has long been recognized. These represent sequential changes in metamorphic facies, ranging from a low-grade facies to a high-grade facies in any one area. Miyashiro (1961, 1973) suggested that these changes in metamorphic facies should be designated metamorphic facies series, and he recognized four major facies series with two

intermediate “groups.” In order of increasing prevailing pressure these are as follows.

1. Contact (thermal) metamorphic facies series
2. Andalusite-sillimanite facies series (*Abukuma type*)
3. Low pressure intermediate group (*Buchan type*)
4. Kyanite-sillimanite facies series (*Barrovian type*)
5. High pressure intermediate group (*Sanbagawa type*)
6. Jadeite-glaucophane facies series (*Alpine type*)

The use of type areas for such facies series has been adopted by many workers, and Hietanen (1967) extends the number of types of facies series to include Pyrenean and Idahoan types between Buchan and Barrovian, and a Saxonian type between Barrovian and Sanbagawa types (Fig. 1). A facies series is thus an indication of the geothermal gradient pertaining at the time of the metamorphism. The use of the concept of facies series will continue to be useful, but the rigorous definition of various types is now no longer advocated, since slightly different geothermal gradients have been deduced from practically all areas that have been investigated in detail (Fig. 2).

One of the most important features of metamorphism that has frequently been ignored is the polyphase nature of most regional metamorphic rocks. Many of the “type” areas for regional metamorphism have been only cursorily studied in this respect. For some years the difference between Buchan and Barrovian types was ascribed to the relative age of metamorphism and deformation, the Barrovian type being regarded as syntectonic crystallization and the Buchan type as post-tectonic crystallization. However, Johnson (1983) showed that both areas have a similar complex history of at least four major fold episodes, with metamorphic recrystallization taking place during and after each episode in both areas. The difference is now regarded as one of overall depth (pressure difference).

The four major series that Miyashiro (1961) originally suggested are, however, associated with different geotectonic environments and can be taken as representative of the range of geologic conditions under which metamorphism will take place. Clearly, there will be transitions from one facies series to another as there are transitions from one metamorphic facies to its neighbors. It is consequently rather academic to erect a nomenclature for large numbers of facies series.

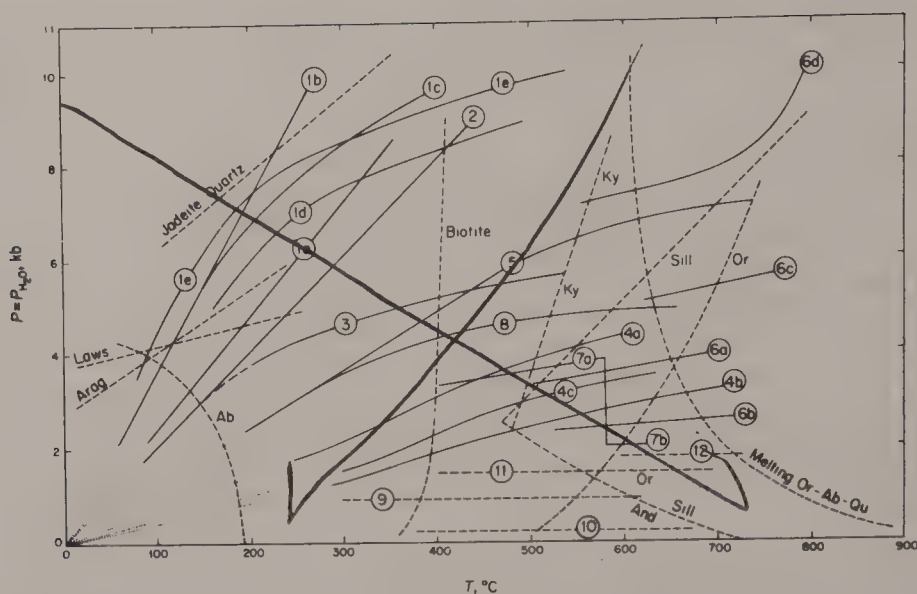


FIGURE 2. Some estimated P - T gradients of individual facies series: (1) glaucophane schist parageneses of northern California (a-c) and Japan (d, e); (2) Eastern Otago schist belt, New Zealand; (3) Alpine schist belt, New Zealand; (4) New Hampshire and Vermont, eastern U.S.A.; (5) southeastern Dalradian (Barrow's zones), Scotland; (6) amphibolite-granulite facies series of northwestern Adirondacks (a), Broken Hill, Australia (b), Saxony (c) and Danube, Austria (d); (7) Nanga Parbat massif and vicinity, Himalayas; (8) Lepontine Alps; (9) Bosost, Pyrenees; (10) Comrie, Scotland; (11) Sierra Nevada, California, U.S.A.; (12) Lochnagar, Scotland. (After Turner, 1968.)

stability of andalusite at lower grade and sillimanite at higher grade in pelitic rocks. Kyanite, chloritoid, staurolite, glaucophane, and lawsonite are absent. Cordierite is commonly present at medium to high grades but almandine is restricted to high-grade pelites. In middle and lower grades, garnet is rare but where present is manganiferous. Synkinematic granites are abundant both in the high-grade zone and outside it.

Greenschist facies (9a), epidote amphibolite facies (10b), amphibolite facies (11a,b,e) and possibly granulite facies (12a,d) assemblages are represented (numbers refer to facies and subfacies as listed under **Metamorphic facies**).

Read (1952) gives an account of Buchan-type assemblages that are intermediate between those of Abukuma and Barrovian types.

Kyanite-sillimanite facies series (Barrovian type). This series is characterized by the stability of kyanite at medium grade and sillimanite at higher grade. Andalusite, cordierite, glaucophane, and lawsonite are absent. Chloritoid, staurolite, and almandine are normal, almandine being abundant in both pelitic and basic rocks from medium grade to high grade. Synorogenic granites often form the core of this type of facies series; granulites do not occur in these cases. Greenschist facies (9a,c), epidote amphibolite facies (10a), amphibolite facies (11d,e,f), and sometimes granulite facies (12b,c,e) assemblages are represented. Eclogites (13, type B) are sometimes found enclosed in synorogenic migmatites. Lower-grade facies (5c, 6f) are thought by McNamara (1965) to be present in the classic area of Dalradian rocks in the Scottish Caledonides but in general this series is characterized by the absence of zeolites and prehnite in the lower grades.

Hashimoto (1972) gives an account of Sanbagawa-type assemblages that are intermediate between those of Barrovian and Alpine types.

Jadeite-glaucophane facies series (Alpine type). This series is characterized by the stability of lawsonite and jadeite-quartz at high grade and by the abundance of zeolites, pumpellyite, and stilpnomelane at low grades. Andalusite, sillimanite, and cordierite do not occur. Kyanite is found at high grade, and pyralspite, both almandine-rich and spessartine-rich, is common. Large volumes of basic and ultrabasic intrusions are characteristic of this type, which may represent upthrust wedges of oceanic crust. Granitic plutons are absent. Zeolite facies (5a,b), prehnite-pumpellyite facies (6a), pumpellyite-actinolite schist facies (7), glaucophane-lawsonite schist facies (8a,b,c,d), and greenschist facies (9c,d) as-

semblages are represented. Eclogites (13, type C) are frequently found in the glaucophane schists.

Geotectonic environment

The main facies series of regional metamorphism are closely controlled by the geotectonic environment. Miyashiro (1961) showed that the jadeite–glaucophane- and the andalusite–sillimanite-types were characteristically arranged in paired orogenic belts, the higher-pressure belt being to the oceanic side and yielding younger ages than the high-temperature belt lying on the continental side of the orogen. This disposition has been interpreted as a consequence of the cordilleran-type of orogeny in which a subduction zone is adjacent to sialic continental crust (Wyllie, 1971). The thick wedge of sediments forming off the continental margin over oceanic crust is dragged down the subduction zone (forming an oceanic trench) and the deeper levels are raised to high pressures, without the normal increase in temperature, owing to the very low geothermal gradient over a subduction zone. This results in the glaucophane–lawsonite schist facies being developed in the sediments. The eclogite fragments and the basic and ultrabasic bodies are regarded as parts of the oceanic crust and upper mantle.

As the oceanic crust is carried down the subduction zones, it reaches depths at which volatiles are expelled into the mantle above, causing partial melting and the uprise of andesitic magmas and calc–alkaline plutons in the continental area on the stable side of the subduction zone. This area then becomes the site of a continuing regional uprise in temperature to fairly high levels in the crust, giving rise to the regional andalusite–sillimanite facies series.

The kyanite–sillimanite facies series, on the other hand, typically develops during a collision-type orogeny, when a continental mass on the active plate is brought next to a continent on the passive side of a subduction zone. The two continental masses are both involved in the resultant very extensive reworking of the sialic crust. Large volumes of continental margin sediments may be depressed deep in the crust for long periods, so that normal geothermal gradients are established and both high temperature and high pressure result; the pelitic rocks develop kyanite- and sillimanite-bearing assemblages. Because of the great volume of hydrous rocks normally involved the higher grades usually exceed the minimum melting temperature for granite under hydrous conditions and partial to complete melting in the highest

grades is common. This results in the development of migmatitic granites and gneisses over large areas. In Precambrian times, many orogenic belts apparently developed without the thick sedimentary sequences and the metamorphism affected large parts of already crystalline rocks. These frequently show upper amphibolite and granulite facies mineral assemblages but without any large scale development of synorogenic granites.

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Cross-references: *Facies of thermal metamorphism; Metamorphic facies; Paired metamorphic belts; Regional metamorphism.*

METAMORPHIC GRADE, METAMORPHIC ZONE

Rocks originating under the same metamorphic conditions are said to be of the same *metamorphic grade*, and planes of equal metamorphic grade are called isograds. In regional and contact metamorphism these are normally boundaries marked by the appearance of an index mineral or mineral assemblage in rocks of a given composition (e.g., chlorite, biotite, garnet in regional metamorphism): the rocks occurring between isograds are referred to as constituting a *metamorphic zone* (e.g., biotite zone between the biotite and garnet isograds). Isograd rocks are those that have originated under closely similar conditions of temperature and pressure (Tilley, 1924). This concept thus expresses in a general way the physical implications of progressive zoning with regard

to temperature, pressure, and other variables. It is thus of much wider use than the term "metamorphic facies," which only refers to mineralogical content.

The arrangement of metamorphic zones of increasing metamorphic grade may produce a distinct overall pattern, e.g., a thermal anticline or thermal dome.

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Cross-references: *Aureole*; *Contact (thermal) metamorphism*; *Metamorphic facies*; *Regional metamorphism*.

METAMORPHIC MINERAL GROWTH

During metamorphism, sedimentary or igneous assemblages of minerals become unstable as a result of newly-imposed conditions of temperature and pressure, and the minerals interact to form new assemblages. Hence, crystal growth in metamorphic rocks takes place at temperatures of ca. 200–800°C, and at pressures ranging up to 8–10 kb, corresponding to a depth of approximately 30 km. Since crystals of mica and other minerals commonly have a parallel orientation, some crystal growth may take place under conditions of directed pressure.

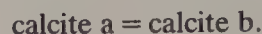
Crystal growth in metamorphic rocks is interpreted from (1) the size, shape, orientation, and distribution of crystals (i.e., rock textures), and (2) the experimental study of the deformation and recrystallization of minerals and rocks at elevated temperatures and pressures.

Although crystal growth in metamorphic rocks is presumably a very slow process, each newly-formed crystal must have passed through a stage of nucleation, followed by a period of enlargement or growth. Thus, there is no reason to suppose that the process is fundamentally any different from crystal growth in a solution or vapor phase at low pressures (see Nancollas and Purdie, 1966).

Classes of mineral growth

It is possible to recognize two general classes of crystal growth in metamorphic rocks, based on inferences concerning the nature of the chemical reactions that have taken place to produce the crystals.

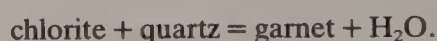
Recrystallization. During metamorphism, changes commonly take place whereby some crystals of a mineral species grow at the expense of others of the same species, a process generally known as recrystallization. This may be regarded as a chemical reaction, which for the recrystallization of calcite may be written



Reactions of this kind have been reproduced experimentally (Griggs et al., 1958). The growth of crystals during recrystallization appears to take place as a result of the jumping of atoms across interfaces between crystals, thus allowing some grains to grow while others experience dissolution.

Two forms of energy may be responsible for these reactions: (1) strain energy, which is reduced as strain-free crystals grow at the expense of strained crystals, and (2) interfacial energy, which is reduced as the mean crystal size is increased, and as the interfaces assume certain patterns. The pattern may be, for instance, a mosaic with triple grain junctions and grain-boundary angles close to 120°. Although recrystallization is very common in nature, it is often difficult to identify its cause.

Crystallization of new minerals. Many chemical reactions that take place in metamorphic rocks produce new minerals at the expense of old minerals. Thus, crystals of garnet, staurolite, kyanite, and many others are produced by reactions of which the following forms an example:



Nucleation is an essential step in most or all of these reactions. The microscopic examination of metamorphic rocks often reveals, however, that the product crystals have not necessarily nucleated and grown near the reactant crystals, and a rather complex rearrangement of matter must take place as the reaction progresses. Also, the growth of the newly formed crystal must be accompanied by the removal of other mineral grains from the path of crystal growth, possibly by chemical dissolution of these crystals and migration of the constituent atoms to other places in the rock.

Reactions of the kind presently considered obviously proceed because the products have a lower bulk free energy than the reactants. Although many reactions of this kind have been reproduced experimentally, no significant progress has been made to date in the simulation of corresponding rock textures by experimental methods.

Growth textures

Some aspects of the textures of metamorphic rocks provide information concerning the nucleation and growth of crystals within them (Harker, 1939; Ramberg, 1952; Spry, 1969).

Crystal shape. Some metamorphic minerals (e.g., staurolite, andalusite) commonly occur as euhedral (idiomorphic) crystals, i.e., crystals that impose their form on the surrounding minerals, while others (e.g., feldspar, calcite) almost invariably occur as anhedral (xenomorphic) crystals. Forms that develop in these rocks are invariably simple forms of low crystallographic indices, and the faces are commonly parallel to cleavage planes. The presence of euhedral forms suggests that the crystals have grown by the development of layers on the crystal faces, and the spreading of these layers to the edges of the faces. Hence the steps observed on the faces of some crystals in marble, notably phlogopite and graphite, may be growth steps, analogous to those observed on crystals grown from the vapor phase. Since crystals in metamorphic rocks presumably grow very slowly, under conditions of low "supersaturation," it is not surprising that complex forms, with faces of relatively high interfacial energy are generally absent.

With regard to nonequidimensional crystals such as mica and amphibole, a great variation may be found in the ratio of dimensions, e.g., the ratio of diameter to height of prismatic phlogopite crystals in marble may range from about 3 or more in some rocks to about 0.5 in others, and the ratio may vary considerably within a small volume of rock. These variations are likely to be the result of variation in the growth rate of "horizontal" faces relative to the "vertical" faces of the prismatic crystals, and are ultimately due to a number of factors, including the degree of "supersaturation," the chemical composition of the nutrient "dispersed phase," and the nature of the crystalline material lying in the path of crystal growth.

Crystal orientation. One of the most striking attributes of metamorphic rocks is the common presence of a foliation and lineation, defined by the parallel or near-parallel arrangement of platy and elongate crystals. The orientation of these planar and linear features may often be related to axial planes and axes of folds, or to other structures, and although the cause of crystal orientation is not yet fully understood, it is undoubtedly related in some manner to rock deformation. It seems likely that in many rocks, crystal nucleation and growth take place while the rocks are experiencing a large-magnitude, penetrative deformation (Turner and Weiss, 1963; Flinn, 1965; Vernon, 1976).

Crystal size. The size of crystals in most metamorphic rocks falls within the range 0.1–10 mm, but some low-grade rocks and mylonites are very fine-grained, and some skarns contain crystals that are several centimeters in dimension. Data presently available on the crystal-size distribution for some minerals, notably garnet, in small volumes of rock indicate that a normal or positively-skewed size distribution is common. Information of this kind may be interpreted in relation to the rate of nucleation. Thus, it may be supposed that the relatively few large crystals represent the initial stage of crystallization, when the rate of nucleation was low: the many medium-sized crystals represent the subsequent, more productive stage, when the rate of nucleation was high, and the few small crystals represent the final stage of crystallization, when the rate of nucleation was declining.

Inclusions. Relatively large crystals in metamorphic rocks commonly contain inclusions of other minerals, and the disposition of these may provide information on the growth history of the host crystal. For example, the distribution of small foreign inclusions in andalusite crystals, to produce a cross-like pattern, would seem to indicate that foreign material was expelled from the faces as they advanced during growth, but not from the edges of the crystals. Also, the arrangement of inclusions in garnet crystals, in the form of a sigmoid (Fig. 1), suggests that the crystals experienced rotation relative to the matrix during growth (Fig. 2).

Spatial distribution. Some information concerning nucleation sites may be obtained by examining the spatial distribution of crystals in rocks. Thus, the presence of reaction rims (e.g., rims of amphibole and quartz about crystals of calcic pyroxene) indicates that nucleation of new phases occurred near the initial reaction interface. More commonly, however, the product crystals are found scattered throughout the rock, intermingled with crystals of other minerals, presumably because the rate of increase of temperature and the rate of reaction were sufficiently low to permit nucleation of the products to be a highly selective process. This may explain, for example, the location of sillimanite crystals at various points in a rock, not necessarily near andalusite crystals that evidently acted as a source for the sillimanite.

Some minerals, notably the amphiboles, may form radiating aggregates, evidently the result of selective nucleation; in general aggregates or clusters of like crystals are rather common. However, crystals of garnet and other minerals in some rocks are randomly distributed, and a more dispersed, or regular distribution exists in some rocks. Many of these distribution patterns

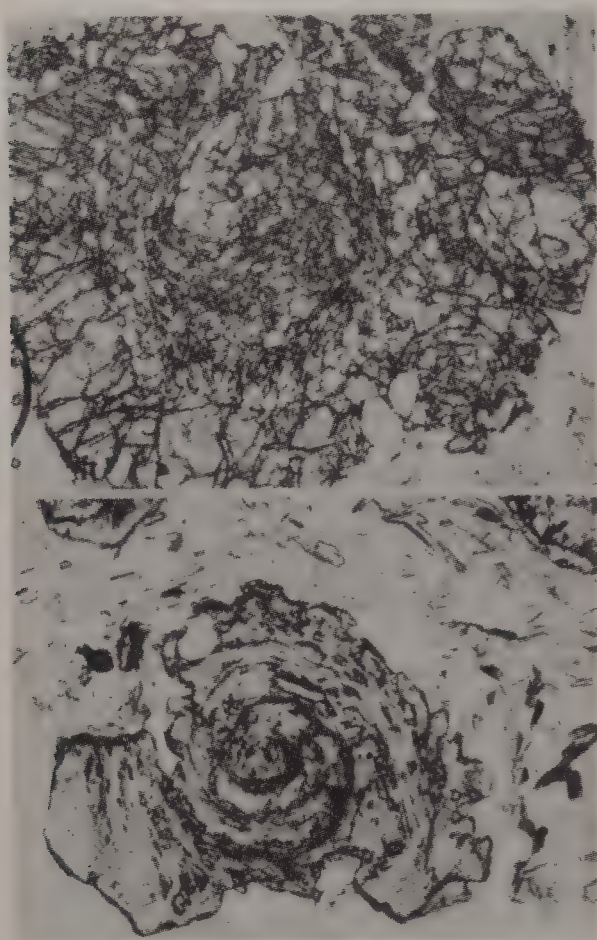


FIGURE 1. Garnet crystal with inclusions in the central portion defining a sigmoidal pattern; crystal diameter is 4 mm. (From Spry, 1969.)

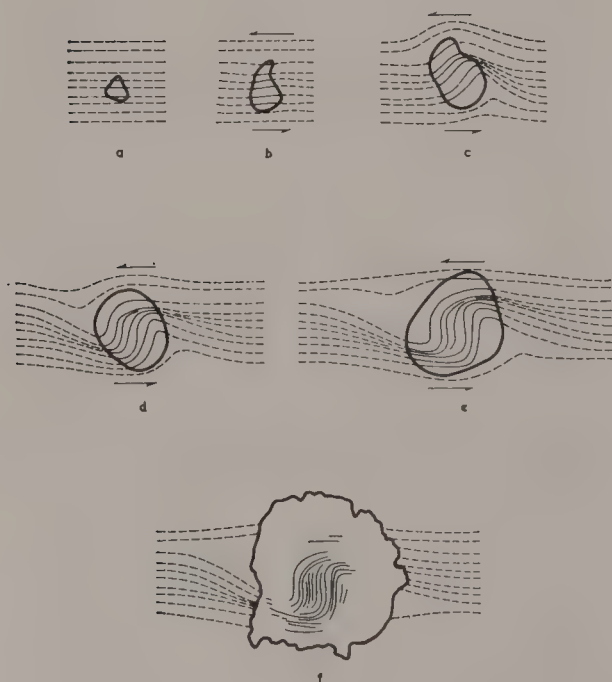


FIGURE 2. Reconstructed stages in the growth of the garnet crystal of Fig. 1. (From Spry, 1969.)

may be determined by the abundance and distribution of energetically favorable nucleation sites.

Kinetics of crystallization

Certain rocks may yield information on the kinetics of the crystallization processes that occurred within them. Thus, certain metamorphic rocks that contain newly formed crystals of garnet that are (1) isolated from each other by a matrix of other minerals, and (2) compositionally zoned, may be suitable for study (Kretz, 1973).

The first step in the analysis is to examine compositional profiles in crystals of different sizes from a small volume of rock and, from these results, to deduce an equation for the rate of crystal growth. For example, if the crystals grew according to the relation $dr/dt = k$, where r , t , and k are crystal radius, time, and rate constant, respectively, then a particular increment of concentration of an element (e.g., Fe) in garnet should correspond to a constant increment of crystal radius, regardless of the size of the crystal being examined. If, on the other hand, the growth equation was $da/dt = k$ or $dv/dt = k$, where a and v are crystal area and volume, a particular increment of concentration will correspond to an increment of radius that will not be constant and will depend on crystal size.

The next step is to determine the crystal-size distribution and obtain therefrom an expression for the rate of nucleation, dn/dt , where n is the number of crystals present in unit volume of rock. Finally, the expressions for the rate of growth and the rate of nucleation may be combined to obtain an equation for the volume, V of garnet in unit volume of rock as a function of time:

$$V(t) = \int_{u=0}^{u=t} v(t, u) \left[\frac{dn}{dt} \right]_{t=u} du$$

where $v(t, u)$ is the volume of a crystal that began growth at time u when observed at time t .

In one garnet-bearing rock that crystallized in a broad contact aureole and was examined from this point of view, it was found that the growth rate was such that (1) the crystal radius was increasing at a nearly constant rate (i.e., the rate of increase of volume was a function of surface area, and growth was controlled by the placement of atoms on the surface rather than the diffusion of atoms to the surface), (2) the nucleation rate was acceleratory, following approximately a power-law expression, and (3) the rate of production of garnet (i.e., the crystallization rate or the reaction rate) was highly acceleratory. The garnet-forming reac-

tion evidently began slowly, then increased rapidly, and finally terminated abruptly, presumably when one of the reactant minerals was consumed. A near-linear rise in temperature could account for the observed relationships.

R. KRETZ

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Cross-references: *Metamorphic petrology—use of thermodynamics; Nucleation of metamorphic minerals; Polyphase metamorphic mineral growth; Reaction rates in metamorphism.*

METAMORPHIC PETROLOGY—USE OF THERMODYNAMICS

An increasing number of studies of metamorphic petrology are leaning, to a lesser or greater extent, on the theories of classical chemical thermodynamics. Thermodynamics may be used (1) to calculate pressure–temperature–fluid composition conditions of metamorphism of observed assemblages, (2) as a theoretical base for discussing phase relations and compositions within a specific system, or (3) as a framework within which diffusion and reaction kinetics may be discussed. Of these, emphasis here is on (1), although the others are at least as important.

The relationship between the composition of phases, both solid and fluid, that constitute a metamorphic assemblage and the pressure and temperature conditions of metamorphism is

necessarily complex. However, the use of equilibrium thermodynamics permits the calculation of values for P – T – X_n that are uniquely limited by the compositions of the coexisting solid phases. This calculation is achieved by modeling an equilibrium system, normally defined by a series of multivariant reactions, in terms of the thermodynamic parameters G (Gibbs free energy), H (enthalpy), S (entropy) and V (volume), and solving for the required unknowns (a list of symbols used is given at the end of the article).

Thermodynamic relationships

As a background to showing how equilibrium thermodynamics may be used in evaluating conditions of formation of metamorphic rocks, the following is a brief introduction to some thermodynamic parameters and their relationships. The chemical potential μ is a measure analogous to electrical or gravitational potential. By definition, a phase or assemblage of low μ is more stable than one of high μ and consequently the flow of matter is always from a region of high μ to one of low μ . Take as an example a reaction between two phases in the Al_2SiO_5 system such that



An equation may be derived such that

$$\Delta\mu = \mu_{\text{Al}_2\text{SiO}_5}^{\text{sillimanite}} - \mu_{\text{Al}_2\text{SiO}_5}^{\text{kyanite}}. \quad (2)$$

This equation fits the general form

$$\Delta\mu = \mu^{\text{products}} - \mu^{\text{reactants}} \quad (3)$$

At point A in Fig. 1 kyanite is stable and any sillimanite present would react to kyanite by transfer of Al_2SiO_5 down a chemical potential gradient, stabilizing the phase of lower μ . At point C the opposite would be the case with sillimanite being the stable low μ phase. At B both phases will be in equilibrium so there

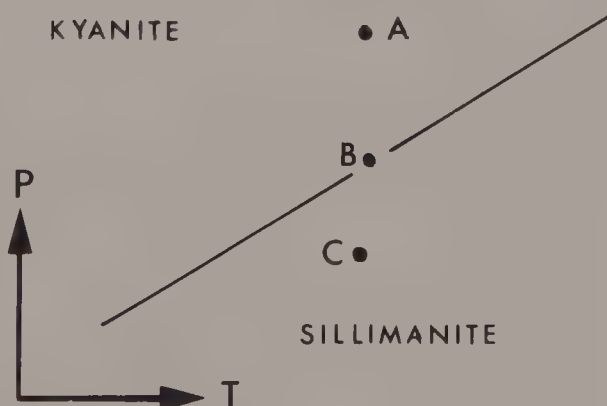


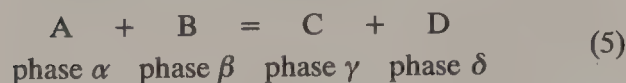
FIGURE 1. General form of the boundary between kyanite and sillimanite in the Al_2SiO_5 system.

should be no transfer of material. Three relationships can be derived.

$$\begin{aligned} \text{at A} \quad \Delta\mu &> 0 \\ \text{at B} \quad \Delta\mu &= 0 \\ \text{at C} \quad \Delta\mu &< 0. \end{aligned} \quad (4)$$

It is clear that μ is a variable that is pressure- and temperature-dependent.

Consider the further reaction



where A, B, C, and D are components of the solid solutions α , β , γ , and δ , respectively, in the way that Mg_2SiO_4 (forsterite) is a component of the solid-solution-phase olivine. The chemical potential of component A in phase α may be written as μ_A^α and, for the olivine example, as $\mu_{\text{Mg}_2\text{SiO}_4}^{\text{ol}}$. At equilibrium, the condition applies that

$$\mu_C^\gamma + \mu_D^\delta = \mu_A^\alpha + \mu_B^\beta \quad (6)$$

(cf. eqn. (3)).

Two important relationships must be introduced here. The first states that the Gibbs free energy of a phase or system is equal to the sum of all the chemical potentials each multiplied by the number of moles of the respective component. For a one component system

$$G = N \cdot \mu \quad (7)$$

while the general case for a multicomponent system is

$$G = \sum_i N_i \mu_i. \quad (8)$$

The second relationship describes the activity of a component i in a phase j as reflecting the difference between the observed chemical potential and that of the standard state:

$$\mu_i^j = \mu_i^\circ + RT \ln a_i^j \quad (9)$$

where μ_i° is the chemical potential of i in the standard state, taken here to be the standard state of the pure phase at the P , T of interest, and a is the activity of component i in phase j , and measures the amount that phase j deviates from the composition of the pure end-member component i such that for a pure phase $a = 1$ and $RT \ln a = 0$.

Rearranging eqn. (6) and substituting eqn. (8) into (9) we derive

$$\begin{aligned} \mu_C^\gamma + \mu_D^\delta - \mu_A^\alpha - \mu_B^\beta &= (\Delta G_r)_{PT} = (\Delta G_r^\circ)_{PT} \\ &+ RT \ln \frac{a_C^\gamma \cdot a_D^\delta}{a_A^\alpha \cdot a_B^\beta}. \end{aligned} \quad (10)$$

The term ΔG_r° is the specific term for the change

in Gibbs free energy during a reaction involving standard-state phases. For any individual phase at standard state, the relation applies that

$$G^\circ = H^\circ - TS^\circ + PV^\circ \quad (11)$$

where H is the enthalpy (heat of formation) of the phase, S is the entropy (effectively a measure of the ordering of the phase), and V is the volume. For a reaction we can write

$$\Delta G_r^\circ = \sum G_{\text{products}}^\circ - \sum G_{\text{reactants}}^\circ \quad (12)$$

which, at equilibrium, gives

$$\begin{aligned} \Delta G_r^\circ &= \Delta H_r^\circ - T \Delta S_r^\circ \\ &+ P \Delta V_r^\circ = -RT \ln K \end{aligned} \quad (13)$$

where $-RT \ln K$ is shorthand for the term on the right-hand side of eqn. (10). In the general form,

$$\ln K = \ln \frac{a_{\text{products}}}{a_{\text{reactants}}} \quad (14)$$

The activity a is related to the phase composition by the relationship

$$a_i^j = (X_i^j \cdot \gamma_i^j)^n \quad (15)$$

where X_i^j is the mole fraction of component i in phase j , γ_i^j is the activity coefficient, and n is the number of sites on which mixing takes place. The activity of a phase is given by the product of the activities on each site. For olivine,

$$a_{\text{Mg}_2\text{SiO}_4}^{\text{ol}} = (X_{\text{Mg}_2\text{SiO}_4}^{\text{ol}} \cdot \gamma_{\text{Mg}_2\text{SiO}_4}^{\text{ol}})^2 \quad (16)$$

where

$$X_{\text{Mg}_2\text{SiO}_4}^{\text{ol}} = \frac{\text{Mg}}{\text{Mg} + \text{Fe}}.$$

For the orthopyroxene $\text{MgAl}_2\text{SiO}_6$, where Al can enter sites M1 and M2 as well as the Z site,

$$a_{\text{MgAl}_2\text{SiO}_6}^{\text{opx}} = (X_{\text{Al}}^{\text{M1}} \cdot \gamma_{\text{Al}}^{\text{M1}})(X_{\text{Mg}}^{\text{M2}} \cdot \gamma_{\text{Mg}}^{\text{M2}}) \quad (17)$$

where

$$X_{\text{Mg}}^{\text{M2}} = \frac{\text{Mg}}{\text{Mg} + \text{Fe}} \text{ on site M2,}$$

and

$$X_{\text{Al}}^{\text{M1}} = \frac{\text{Al}}{\text{Al} + \text{Mg} + \text{Fe}} \text{ on site M1.}$$

The activity coefficient (γ) introduced in eqn. (15) takes into account the deviation from ideality of on-site mixing. Mixing models involve solid solutions of two or more components and assume random mixing of large numbers of atoms on each site. Solid solutions that mix ideally have $\gamma = 1$ and $a = X$. With increasing nonideality, γ deviates from unity and $a \neq X$.

An ideal solution can be defined as one in which there is no contribution to the free energy through an additional heat of mixing term in the enthalpy value. In this case, mixing is complete at all temperatures. Where there is a positive heat of mixing contribution, a solvus will be present in the system with near ideality only approximated to at temperatures above that of the solvus crest. It is this increase in ideality with increasing temperature that persuades many metamorphic petrologists to view some solid solutions, particularly the significant Fe–Mg binary solid solution, as ideal at temperatures in excess of 500°C.

If a symmetrical solvus is assumed, then γ values can be derived from nonideal binary or ternary solutions, respectively, such that

$$RT \ln \gamma_1 = X_2^2(W_{12}) \quad (18)$$

$$RT \ln \gamma_1 = X_2^2(W_{12}) + X_3^2(W_{13}) + X_2X_3(W_{12} + W_{13} - W_{23}) \quad (19)$$

where subscripts 1, 2, 3 refer to the component in the solid-solution series. W is an interaction parameter between two components that may be experimentally determined, as has been done for garnets. Interestingly, Fe and Mg are found to mix ideally in garnet, while other relationships in the Fe–Mg–Ca–Mn series are nonideal with often substantial values for W . Values for γ have been determined experimentally for plagioclase. The above equations assume a symmetrical solvus, which may not be too invalid. Alternative models and the validity of the symmetrical one are discussed by Powell (1974).

For a gas or fluid, taking a standard state at P° , T , the activity at P bars and T is given by

$$RT \ln a = \int_{P^\circ}^P V dP \quad (20)$$

where, by definition,

$$\int_{P^\circ}^P V dP = RT \ln \left(\frac{f}{f^\circ} \right) \quad (21)$$

where f and f° are the fugacities (thermodynamic pressure) of the gas at P and P° , respectively. The fugacity of a gas (i) is related to the pressure on the gas by a fugacity coefficient Γ by

$$P_i = f_i \cdot \Gamma_i. \quad (22)$$

Two equations may be used to summarize fluid behavior in metamorphic rocks:

$$P_i = P_{\text{fl}} \cdot X_i \quad (23)$$

$$P_{\text{fl}} = P_{\text{H}_2\text{O}} + P_{\text{CO}_2} + P_{\text{CO}} + P_{\text{H}_2} \quad (24)$$

where P_{fl} may equal P_{total} . Values for f and Γ for

H₂O and CO₂ may be calculated using the formulae listed by Powell and Holland (1985, table 2).

Data sets

The estimation of pressure, temperature, and fluid composition conditions of metamorphism by the thermodynamic method involves solution of the equilibrium relation for a balanced reaction:

$$\Delta G_r^\circ + RT \ln K = 0 = \Delta G_r. \quad (25)$$

It is clear that such calculations can only be carried out if sufficient thermochemical data are available in terms of G , H , S , and V . Robie and Waldbaum (1968) listed values for V° , H_f° , S° at a standard state of 1 bar, 298 K, as well as values calculated from heat-capacity equations at a series of higher temperatures. But this data set has been largely superseded by that of Helgeson et al. (1978), which attempts to evaluate the enormous amount of data now available from calorimetric determination of enthalpies and of entropies of individual phases and the experimental determination of P , T positions of reaction equilibria. Helgeson et al. (1978) list values of G° , H_f° , S° , and V at 1 bar 298 K, as well as values for a , b , and c for the heat capacity equation

$$C_p = a + bT + \frac{c}{T^2}. \quad (26)$$

The change in H and S with temperature is obtained by evaluating the integrals

$$H_T - H_{298} = \int_{298}^T C_p dT \quad (27)$$

$$S_T - S_{298} = \int_{298}^T \frac{C_p}{T} dT \quad (28)$$

respectively, which give

$$H_T = H_{298} + \left[aT + \frac{bT^2}{2} - \frac{c}{T} \right]_{298}^T \quad (29)$$

$$S_T = S_{298} + \left[a \ln T + bT - \frac{c}{2T^2} \right]_{298}^T \quad (30)$$

Substituting eqns. (27) and (28), the free energy of a phase at P , T of interest is given by

$$G_{P,T}^\circ = H_{f,298}^\circ + (H_T - H_{298}) - T(S_{1,298}^\circ + (S_T - S_{298})) + (P - 1)V^\circ. \quad (31)$$

Entropies may be derived by integration of heat-capacity data (after eqn. (29)) over a range of temperatures from 0 K upwards. Since such data can be obtained accurately by experiment, tabulated values of entropy can be considered

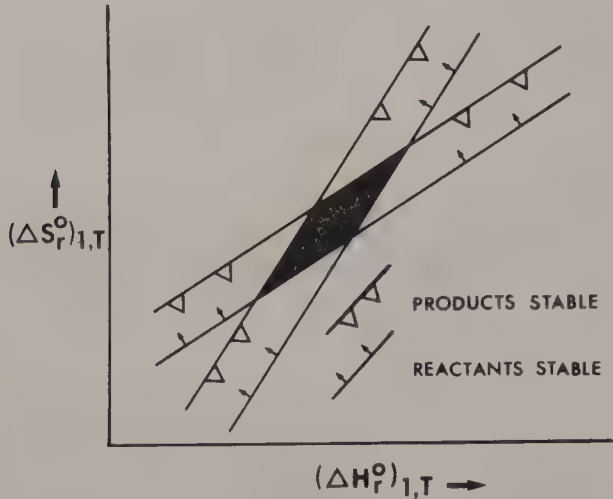


FIGURE 2. Gordon's method: the black area, defined by the intersection of lines in $(\Delta H_r^{(0)})_{1,T}$ – $(\Delta S_r^{(0)})_{1,T}$ space, is the region of all possible consistent values for $(\Delta H_r^{(0)})_{1,T}$ and $(\Delta S_r^{(0)})_{1,T}$.

reliable to within 0.2%, although disordered mixing may have a significant effect on phase entropy. Enthalpy data are less reliable and tabulated data are often inconsistent. Heats of formation have been determined calorimetrically for a number of phases. Modern data derived in such a fashion are extremely reliable, but far more are needed. In the light of such shortcomings, possibly the best method for deriving a thermochemical data set is that outlined by Gordon (1973). It involves the simultaneous solving for a single experimentally determined reaction of a set of inequality

equations, which may be done graphically as shown in Fig. 2. The method produces values for ΔH_r° and ΔS_r° , at a standard state of 1 bar and temperature T , that are derived directly from experimental data. Wells (1976) has shown the inherent assumption of the method, that the enthalpies and entropies are not a function of temperature, to be valid as long as the results are not applied more than a few hundred degrees from the experimental temperatures. The method is elegant, for, as long as the P, T of interest is not far removed from the experimental conditions, the standard state ΔG_r° can be taken at the P, T of interest as

$$(\Delta G_r^\circ)_{P,T} = (\Delta H_r^\circ)_{1,T} - T(\Delta S_r^\circ)_{1,T} + (P - 1)(\Delta V_r^\circ)_{1,298} \tag{32}$$

(Note that to all intents and purposes ΔV_r° at the P, T of interest is equal to that of the 1 bar, 298 K standard state.) An advantage of this method is that entropies and enthalpies need not be directly determined for any particular phase. That ΔS_r and ΔH_r are linearly related is obvious from Fig. 2 and a sensible value of ΔS_r , taken from tabulated data, will yield a consistent value for ΔH_r . A comparison of entropy and free-energy values derived by this method with those calculated from the Robie and Waldbaum (1968) data set is given by Gordon (1973, table 5). While there is good agreement on entropy values, those for ΔG_r are seriously inconsistent, probably reflecting poor-quality enthalpy data. Some values derived by this method are given in Table 1.

TABLE 1. Thermodynamic parameters for some reactions (to fit $\Delta G_r = \Delta H_r - T \Delta S_r + P \Delta V_r = 0$)

musc + qtz = sill + orth + H ₂ O	22,841 – 39.5T – 0.1073P	(1) G
musc + qtz = and + orth + H ₂ O	21,802 – 38.57T – 0.0683P	(1, 2) G
2 phlog + 6 musc + 15 qtz = 3 cord + 8 orth + 6.5H ₂ O	204,070 – 333.58T + 2.6443P	(3) G
2 staur + 7 qtz = 2 cord + 5 sill	33,650 – 46.65T + 2.8217P	(4) G
3Fe-cord = 2 alm + 4 sill + 5 qtz + 1.5H ₂ O	33,325 – 43.95T – 4.01P	(4, 5) G
3Fe-cord = 2 alm + 4 sill + 5 qtz	38,250 – 22.18T – 3.924P	(6) G
3 anorth = gross + 2 kyan + qtz	–11,676 + 32.82T – 1.3006P	(7, 8) G
gross + qtz = anorth + 2 woll	15,574 – 19.6T + 0.7806P	(7, 9, 10) G
4 zois + qtz = 5 anorth + gross + 2H ₂ O	59,800 – 106T + 1.4761P	(9, 10) G
alm + gross + musc = 3 anorth + ann	4,124 – 22.06T + 1.802P	(11)
ann + pyr = phlog + alm	–12,454 + 4.662T – 0.057P	(12)
3 staur + 12.5 qtz = 4 alm + 23 ky + 6H ₂ O	174,657 – 260.97T – 3.699P	(13)
3 staur + 12.5 qtz = 4 alm + 23 sill + 6H ₂ O	212,510 – 325.47T – 0.505P	(13)

G—by Gordon's method.
(1) Evans, B. W. (1965) *Am. J. Sci.* **263**, 647–667; (2) Holdaway, M. J. (1971) *Am. J. Sci.* **271**, 97–131; (3) Seifert, F. (1976) *Contrib. Mineral. Petrol.* **57**, 179–185; (4) Richardson, S. W. (1968) *J. Petrol.* **9**, 467–488—staurolite Fe₄Al₁₈Si₈O₄₆(OH)₂; (5) Holdaway, M. J. and S. M. Lee (1977) *Contrib. Mineral. Petrol.* **63**, 175–198; (6) Currie, K. L. (1971) *Contrib. Mineral. Petrol.* **33**, 215–226; (7) Hays, J. F. (1967) *Carnegie Inst. Washington Yearb.* **65**, 234–239; (8) Hariya, Y. and G. C. Kennedy (1968) *Am. J. Sci.* **256**, 193–203; (9) Newton, R. C. (1966) *Am. J. Sci.* **264**, 204–222; (10) Boettcher, A. L. (1970) *J. Petrol.* **11**, 337–379; (11) Ghent, E. D. and M. Z. Stout (1981) *Contrib. Mineral. Petrol.* **76**, 92–97; (12) Ferry, J. M. and F. S. Spear (1978) *Contrib. Mineral. Petrol.* **66**, 113–117; (13) Yardley, B. W. D. (1981) *Neues Jahrb. Mineral. Mon.* 127–132—staurolite Fe₄Al₁₈Si_{7.5}O₄₄(OH)₄.

By definition, entropies and enthalpies yielded by Gordon's method are an approximation as $\Delta C_p \neq 0$. However, $\Delta S_{f, \text{solids}}^\circ$ (the difference of the entropies of formation of the solids) is temperature-independent and this term has been utilized by Fisher and Zen (1971) and Chatterjee (in Fraser, 1977) in deriving the relationship

$$\Delta G_{r(P,T)} = \Delta H_{298, \text{solids}}^\circ - T \Delta S_{f, 298, \text{solids}}^\circ + (P-1) \Delta V_{298, \text{solids}}^\circ + n G_{\text{H}_2\text{O}(P,T)}^* \quad (33)$$

This equation separates the contributions made to $\Delta G_r(PT)$ by solid and fluid phases, the term $G_{\text{H}_2\text{O}}^*$ being tabulated by Fisher and Zen (1971). Values for $\Delta H_{298, \text{solids}}^\circ$ and $\Delta S_{f, 298, \text{solids}}^\circ$ may be freely evaluated, among other ways by an inequality method similar to that of Gordon. An internally consistent data set derived by this method for reactions involving cordierite–biotite–muscovite–andalusite–sillimanite–orthoclase–quartz in the system KFMASH is given in Droop and Treloar (1981).

Errors enter into the data set through the quality of the raw experimental data. While the errors in P – T calibration inherent in experimental apparatus may be taken into account, a number of other more or less intractable sources of error may be difficult to evaluate. The structural state (ordering) of the starting materials, the type of starting material, the sluggishness or otherwise of a subsolidus reaction, and the production of metastable phases can all affect the quality of the experimental data and hence of the thermodynamic data set derived from it. For example, in deriving the data set mentioned above, Droop and Treloar (1981) showed marked inconsistencies between experiments done in different laboratories. The difficulties involved in determining the position of the Al_2SiO_5 triple point are well known. Experiments on the reaction muscovite + quartz = orthoclase + sillimanite + water show frequent external inconsistencies between individual laboratories. A major problem with subsolidus experiments is the speed of reaction, some reactions being so sluggish that it has proved impossible to bracket the equilibrium temperatures to within $\pm 75^\circ\text{C}$, with the result that the experiments concerned are not particularly useful when it comes to deriving a data set. Data for metamorphic reactions are widely disseminated throughout the literature. Specifically valuable collections include (1) for a variety of medium to high grade pelitic reactions (Hodges and Crowley, 1985); (2) for geothermometers and geobarometers especially relevant to granulite facies assemblages (Essene, 1984, in Ferry, 1984, and Newton 1983); (3) for low-pressure garnet-free contact metamorphic assemblages (Droop and

Treloar, 1981); (4) for geothermometry and barometry in the garnet–biotite–cordierite–sillimanite–quartz assemblage (Treloar, 1981); (5) for reactions in the siliceous dolomite series (Skippen, 1974, but see Slaughter et al., 1975); and (6) for barometry using garnet–rutile–ilmenite assemblages with either plagioclase, kyanite, or sillimanite in many amphibolite and granulite facies rocks (see Bohlen et al., 1983, and Bohlen and Liotta, 1985). Helgeson et al. (1978), as well as providing thermodynamic data for individual phases give an extensive discussion of many experiments in all systems. However, probably the most comprehensively readily utilized data base is THERMOCALC (Powell and Holland, 1988) which the reader is encouraged to use.

Application

Once in possession of a data set and electron–microprobe analyses of the coexisting phases that are contained within a rock, the petrologist is in a position to calculate the metamorphic conditions at which the observed assemblage was at equilibrium. The data set provides thermochemical data for the end-member reactions, which can then be corrected for the specific case by using the analyzed phase compositions to calculate the a – X relationships in the $RT \ln K$ part of eqn. (25). The major variables that are of interest in an evaluation of the conditions of metamorphism are P , T , $X_{\text{H}_2\text{O}}$, and X_{CO_2} . Generally, as many equations must be solved as there are variables of interest, where each equation involves the thermodynamic modeling of an equilibrium reaction.

The ideal case for a pelite would be to use (1) a geothermometer, such as the two-phase garnet–biotite Fe–Mg exchange reaction, to estimate temperature; (2) a solid–solid reaction, preferably with a flat slope in P – T space (low value of $\Delta S/\Delta V$) to estimate pressure at the relevant temperature; and (3) a dehydration reaction to calculate $f_{\text{H}_2\text{O}}$ at the estimated P, T conditions. An oxidation reaction, such as magnetite to hematite, may yield a value for f_{O_2} which, if graphite is present, can be used to estimate f_{CO_2} , f_{CO} , etc.

Unfortunately, the ideal case is not the norm, especially in pelitic rocks where dehydration reactions dominate and solid–solid reactions are unusual. In the likely absence of a barometer, a thermometer, or both, it becomes necessary to solve for two or three variables simultaneously. This is typically so in the case of the siliceous dolomite system, where, if P can be estimated, simultaneous solving of two equations will give values for T and X_{CO_2} at P . Ideally, thermodynamic calculations should be carried

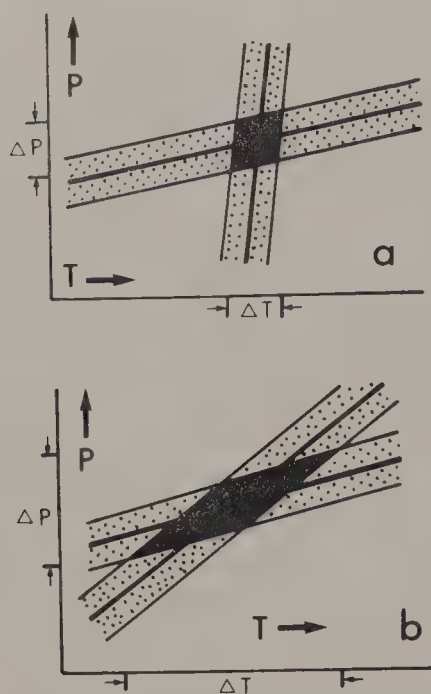
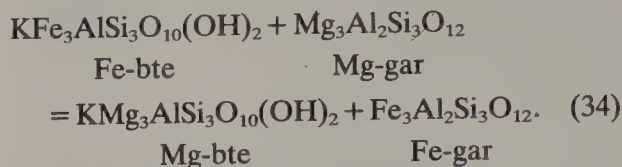


FIGURE 3. P - T diagrams showing the increase of the uncertainties in the calculated temperatures and pressures (shown by black) as the angle of intersection between two reaction lines decreases; the uncertainty in the position of each line is shown by stipple.

out for reactions with a high angle of intersection in the two-dimensional space used. A barometer and a thermometer, for instance, will give reaction lines at about 90 degrees to each other in P - T space (Fig. 3a). As the angle of intersection decreases, the amount of uncertainty in the calculated pressure and temperature increases (Fig. 3b).

The estimation of errors associated with thermodynamic calculations is of importance, and a number of parameters have to be considered. The quality of the experimental data and subsequently derived data set, the accuracy of the microprobe data (which, on an energy-dispersive instrument, is unlikely to be better than $\pm 1\%$), and the validity of activity models used will all have an effect on the final answers. It should be remembered that even on a well-studied experimental exchange reaction thermometer such as biotite-garnet (Ferry and Spear, 1978) the errors are of the order of $\pm 50^\circ\text{C}$ even before problems of substituting other elements are considered. As most dehydration reactions and many solid-solid reactions have steep slopes in P - T space, pressures may often not be verifiable to within 1 kb *at best*. In addition, errors have a habit of propagating throughout a calculation and some attempt to account for this should be made—but rarely is.

Example of geothermometry. The presence of coexisting garnet and biotite in a rock enables temperatures of metamorphism to be calculated by use of the Fe-Mg exchange reaction geothermometer:



This reaction has been experimentally studied by Ferry and Spear (1978) who derive for the reaction the equation

$$-12,454 + 4.662T - 0.057P + 3RT \ln K = 0$$

where

$$K = \frac{\text{Fe-gar} \cdot \text{Mg-bte}}{\text{Mg-gar} \cdot \text{Fe-bte}}$$

Rearranging,

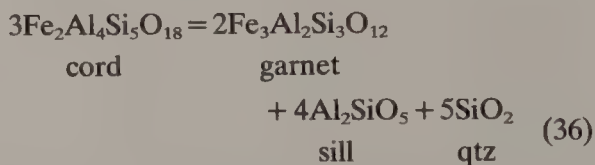
$$T(^{\circ}\text{C}) = \frac{4,151 + 0.019P}{1.554 + R \ln K} - 273^{\circ} \pm 50. \quad (35)$$

For a garnet of composition $\text{Fe}_{2.4}\text{Mg}_{0.6}\text{Al}_2\text{Si}_3\text{O}_{12}$ and a biotite of composition $\text{KFe}_{1.5}\text{Mg}_{1.5}\text{AlSi}_3\text{O}_{10}(\text{OH})_2$

$$K = \frac{0.8 \times 0.5}{0.2 \times 0.5} = 4$$

and at $P = 5$ kb, $T = 712^{\circ}\text{C} \pm 50^{\circ}\text{C}$. (Note that the addition of extra components—Mn, Ca in the garnet; Ti, Fe^{3+} in the biotite—will rapidly invalidate the thermometer, but see the analysis of Hodges and Spear, 1983.)

Example of geobarometry. When all the constituent phases are in equilibrium, the reaction



constitutes a barometer for which

$$\Delta G_r = 38,250 - 22.18T - 3.924P + RT \ln \frac{(a_{\text{gar}})^2}{(a_{\text{cord}})^3} = 0. \quad (37)$$

In the Mg-free system, $\ln K = 0$ as a_{garnet} and a_{cord} are both equal to 1. At $T = 600$ and 850°C , equilibrium pressures of 4.813 and 3.400 kb, respectively, are obtained. Adding Mg to the system so that garnet of composition $\text{Fe}_{2.7}\text{Mg}_{0.3}\text{Al}_2\text{Si}_3\text{O}_{12}$ and cordierite of composi-

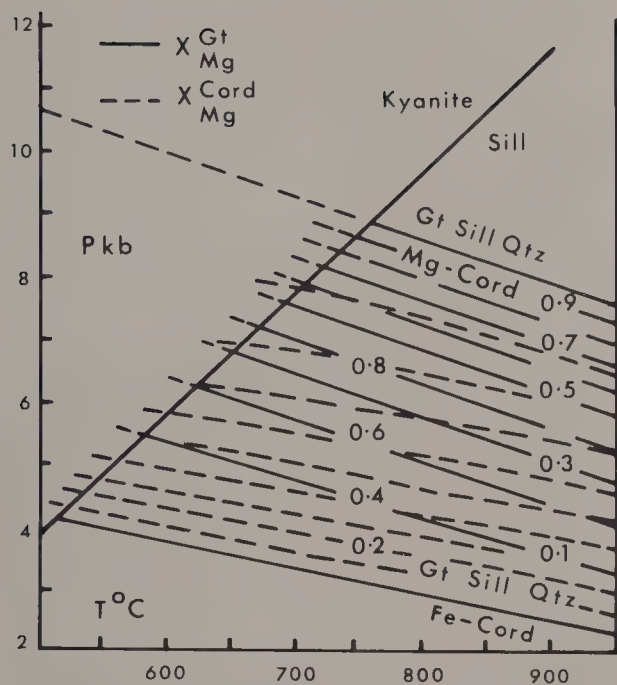


FIGURE 4. The geobarometer cord = gt + sill + Qtz, showing the effect that changing Fe:Mg ratios have on the isothermal equilibrium pressure.

tion $\text{Fe}_{1.5}\text{Mg}_{0.5}\text{Al}_4\text{Si}_5\text{O}_{18}$ are present, then, as $a_{\text{garnet}} = 0.9^3$ and $a_{\text{cord}} = 0.75^2$, pressures of 5.297 kb (600°C) and 4.022 kb (850°C), respectively, are obtained. The equilibrium constitutes a good barometer, as a slight change in the Fe:Mg ratio of the system markedly affects the isothermal equilibrium pressure of the reaction (Fig. 4). The slope of any reaction may be derived from the relationship

$$d(\Delta G) = 0 = \Delta V dP - \Delta S dT \quad (38)$$

which gives

$$\frac{dP}{dT} = \frac{\Delta S}{\Delta V} \quad (39)$$

(the Clausius–Clapeyron equation). For reaction (36), the slope of -5.65 bar/deg is flat, especially when compared with that of the biotite–garnet thermometer (81.80 bar/deg). In reality, as cordierite contains up to $\frac{1}{2}$ mole of water, the amount being dependent on $P_{\text{H}_2\text{O}}$, reaction (36) is a dehydration one and its position in P – T space is in part dependent on varying $X_{\text{H}_2\text{O}}$ in the pore fluid. As a result, it is not as suitable for barometry as at first appears (see Treloar, 1981, and references therein).

Example of fluid-bearing reactions. The effect of changing fluid composition can be shown for the reaction

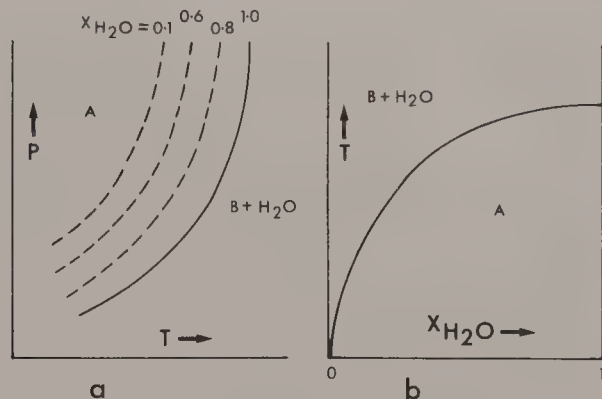
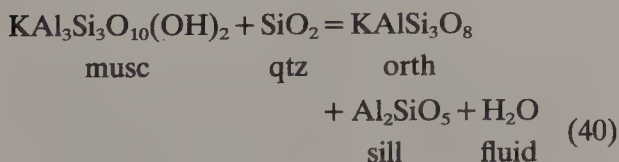


FIGURE 5. The effect of changing values of $X_{\text{H}_2\text{O}}$ on the equilibrium P – T condition for a dehydration reaction: (a) for P – T space; (b) for isobaric T – $X_{\text{H}_2\text{O}}$ space.

for which

$$\begin{aligned} \Delta G_r &= 22,841 - 39.5T - 0.1073P \\ &= RT \ln a_{\text{solids}} - RT \ln a_{\text{H}_2\text{O}}. \end{aligned} \quad (41)$$

With pure solid phases, $a_{\text{solids}} = 1$ and $\ln a_{\text{solids}} = 0$, and at a pressure of 3 kb, the following are obtained:

$$\begin{aligned} T &= 640^\circ\text{C}, & P_{\text{H}_2\text{O}} &= 3,000 \text{ bar} = P_{\text{tot}} \\ T &= 600^\circ\text{C}, & P_{\text{H}_2\text{O}} &= 1,920 \text{ bar} = 0.64P_{\text{tot}} \\ T &= 560^\circ\text{C}, & P_{\text{H}_2\text{O}} &= 1,170 \text{ bar} = 0.39P_{\text{tot}}. \end{aligned}$$

From this, it is clear that the equilibrium temperature of a dehydration reaction increases at constant pressure with increasing $X_{\text{H}_2\text{O}}$ (Fig. 5a). In addition, the slope of a dehydration reaction, such as this one, in isobaric T – $X_{\text{H}_2\text{O}}$

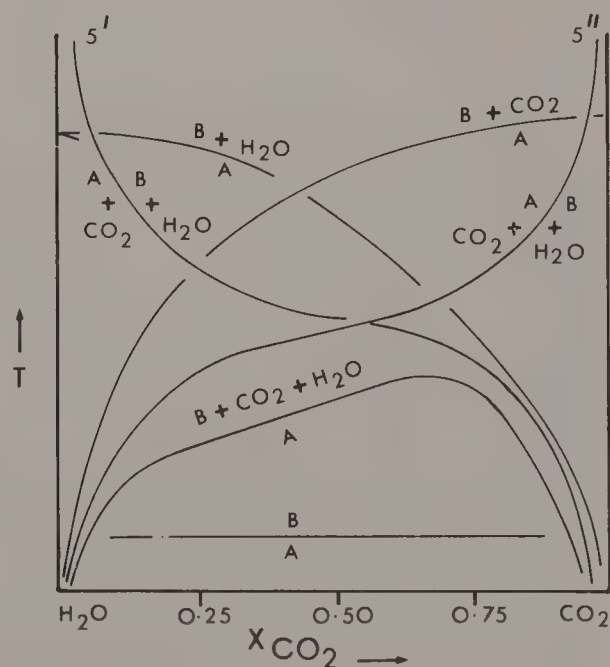


FIGURE 6. T – X_{CO_2} plot of varying types of mixed volatile equilibria: $5' - n_{\text{H}_2\text{O}} > n_{\text{CO}_2}$; $5'' - n_{\text{CO}_2} > n_{\text{H}_2\text{O}}$.

space is concave towards the high- $X_{\text{H}_2\text{O}}$ axis (Fig. 5b). In a closed system, the water released by this reaction with increasing temperature will buffer the fluid phase to increasingly higher values of $X_{\text{H}_2\text{O}}$ along the path in T - $X_{\text{H}_2\text{O}}$ space defined by the reaction. In an open system, the fluid phase will be controlled externally and the reaction will have no significant effect on the fluid phase.

The slope of a mixed volatile reaction (one that involves either H_2O , CO_2 , or both) in H_2O - CO_2 vs. T space is given by

$$\frac{dT}{dX_{\text{CO}_2}} = \frac{RT}{\Delta S_r} \left[\frac{n_{\text{CO}_2}}{X_{\text{CO}_2}} - \frac{n_{\text{H}_2\text{O}}}{X_{\text{H}_2\text{O}}} \right] \quad (42)$$

(Kerrick, 1974), which, for a variety of reaction types, yields Fig. 6 which shows the interdependence of temperature and fluid composition in any fluid-bearing reaction.

Discussion

Reactions (36) and (40) used above as examples illustrate two important points. First, that multivariance in a reaction, while not essential to the thermodynamic method, provides a powerful tool in the estimation of metamorphic conditions. The cordierite-garnet reaction (eqn. (36)) is univariant only in the pure end member FeO - Al_2O_3 - SiO_2 system, while the muscovite-quartz breakdown reaction is univariant only in the system K_2O - Al_2O_3 - SiO_2 - H_2O . The addition of MgO to the former system will make that reaction divariant, and the addition of Na to the latter system will have the same effect. The further addition of CO_2 to the fluid phase in eqn. (40) will have the effect of making that reaction trivariant. It is apparent that multivariance, at least to the di- or trivariant level, must be a common characteristic of metamorphic assemblages. Second, for the thermodynamic method to be valid, phase and fluid compositions must reequilibrate rapidly in response to external stimuli such as a change in temperature. This fundamental assumption underlies the theory of how continuous divariant reactions constantly modify phase compositions. It should be noted that equilibria used in thermodynamic calculations are balanced chemical reactions between chemical components present in the mineral phases in the rocks. It is a common misconception to equate these chemical component equilibria with actual chemical reactions occurring between minerals during metamorphism. Whilst in some cases these may be similar, this is not necessary for equilibria in natural rocks; i.e., the "reactions" used in P - T calculations do not need to occur.

It is important that phases studied are in equilibrium with each other. The common

assumptions that phases may be in equilibrium either directly by contact or indirectly through the medium of the fluid phase seem valid, although the petrologist must beware of the vagaries of domainal or mosaic forms of local equilibrium.

There is far more to the application of thermodynamics in metamorphic petrology than has been discussed in this article. The use of G - X diagrams in constructing solvuses and the G - X basis for deriving a petrogenetic grid are discussed in Powell (1978); the role of μ - μ diagrams in studying metasomatic features and indeed the whole field of nonequilibrium thermodynamics have, of necessity, not been considered here. The theoretical use of the thermodynamic method in the appraisal and evaluation of experimental data has been elegantly illustrated by Helgeson et al. (1978), although the work at the University of Chicago (Newton and Perkins, 1982 and references therein) must not be underestimated. However the cautionary notes of Essene 1984, (in Ferry, 1984) cannot be stressed too highly.

The thermodynamic method, its theoretical basis, and its applications are more fully explored in Kerrick (1974), Wood and Fraser (1976), Fraser (1977), Helgeson et al. (1978), Powell (1978), and Ferry (1984).

PETER J. TRELOAR

List of symbols

G	Gibbs free energy
$G_{\text{H}_2\text{O}}^*$	Particular value for Gibbs free energy of water tabulated by Fisher and Zen (1971)
H	Enthalpy
H_f°	Enthalpy of formation from elements
P	Pressure
P_n	Total fluid pressure
P_{tot}	Total confining pressure
R	Gas constant = 1.987 cal/bar
S	Entropy
S_f°	Entropy of formation from elements
V	Volume
X	Mole fraction
X_n	Composition of the fluid phase
a	Activity
f	Fugacity
$^\circ$	Superscript implying standard state
γ	Activity coefficient
μ	Chemical potential
Γ	Fugacity coefficient
$\Delta G_r^\circ, \Delta H_r^\circ, \Delta S_r^\circ, \Delta V_r^\circ$	Change of G, H, S , and V with reaction at standard state
$\mu_i^j, a_i^j, \gamma_i^j, X_i^j$	μ, a, γ, X of component i in phase j

- P_i, F_i, Γ_i Partial pressure, fugacity, or fugacity coefficient of a gas species i
- $()_{P,T} ()_{1,T} ()_{1,298}$ Subscripts relating to standard state at P, T ; 1 bar, T ; 1 bar, 298°C; of interest
- $()_{\text{solids}}$ Subscript relating to (H, S, V, a) of the solid phases in a reaction

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Cross-references: *Contact (thermal) metamorphism*; *Facies of thermal metamorphism*; *Metamorphic facies*; *Metamorphic mineral growth*; *Metamorphic reactions*; *Metamorphism—controls*; *Regional metamorphism*.

METAMORPHIC REACTIONS

There are a limited number of types of reactions that occur commonly during metamorphism.

Polymorphic transitions

When a substance (element or compound) of given chemical composition can exist in two or more modifications with distinct atomic arrangement, these distinct arrangements are termed polymorphs. It can be shown from the arguments of classical thermodynamics that under randomly selected P – T conditions one form will have the lowest chemical potential and be the stable form. In Fig. 1, experimental data concerning the regions of stability of four common examples of polymorphic modifications found in metamorphic rocks are presented.

Transitions involving the Al_2SiO_5 polymorphs (andalusite, sillimanite, kyanite) are sluggish, and hence these minerals are readily preserved. At very high temperatures, these phases break down to mullite ($\text{Al}_6\text{Si}_2\text{O}_{13}$) and silica. This reaction is seen when some rocks are enclosed in silicate melts at the upper limit of metamorphic temperatures.

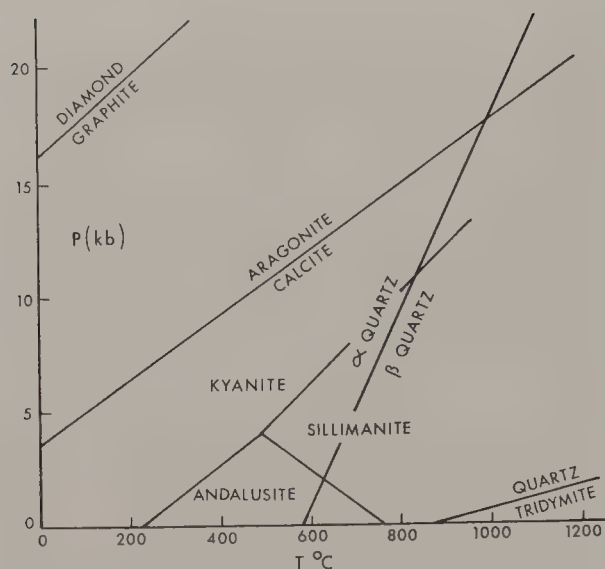


FIGURE 1. Region of stability of several systems of polymorphs.

Calcite is the common form of calcium carbonate in metamorphic rocks, but in rocks crystallized at high pressures aragonite should be stable. In this case, the transition takes place quite rapidly in the solid state at moderate temperatures and it is only in rocks that have been subjected to high pressures at low temperatures that survival of aragonite is observed (e.g., rocks of the glaucophane schist facies).

Almost all carbon in metamorphic rocks occurs as the low pressure polymorph, graphite. In some rocks of the eclogite facies, diamond is found. It is doubtful whether any surface rocks could achieve the diamond field by burial, and diamond-bearing rocks occur as inclusions in igneous rocks that come from a deep-seated mantle environment.

The occurrence of the polymorphic modifications of SiO_2 indicates the caution with which interpretation of physical conditions must be approached. Most metamorphic rocks contain quartz, as would be anticipated from Fig. 1. Some high-temperature members of the sanidin-facies contain tridymite, but in rocks of low temperature (ca. 200°C) hydrothermal metamorphism, cristobalite may be precipitated from highly supersaturated silica solutions.

All the cases involve reconstructive, first-order polymorphic changes where two crystal structures are quite distinct. Such transitions are frequently sluggish and hence high-temperature forms tend to be preserved. If polymorphic transitions are rapid and reversible, as with some first-order and many second-order changes, preservation may not occur. The α - β quartz transition (Fig. 1) is typical of rapid reversible transitions and must occur reversibly in many metamorphic rocks. In favorable cases

this may be deduced by observation of the preservation of some crystal form limited by symmetry to one modification.

Second-order transitions include the Al-Si ordering and disordering processes seen with the alkali feldspars, where the processes are sluggish. In general, the degree of ordering increases with increased pressure and decreases with rise in temperature. Again, metastable products may be formed at low temperatures. In supersaturated solutions in hydrothermal environments, the high-temperature modifications may be produced.

The conditions necessary for stable formation of a given polymorph can normally be measured in the laboratory with reasonable precision. Obviously, they are important physicochemical markers, for their formation depends only on the variables P and T . This simple situation is slightly complicated by significant quantities of impurities. Normally, one crystal structure modification is more tolerant of a certain impurity than another and this modification may be stabilized by the impurity. For example, SrCO_3 will stabilize aragonite in preference to calcite. Few studies have been conducted to assess such effects, but if the laws of ideal solutions are applied it is apparent that the quantities of impurities necessary to cause any drastic shift are of the order of several percent.

Solid-solid transitions

As with polymorphic transitions, solid-solid transitions depend only on P and T . A good example is provided by the stability of jadeite, which may occur in silica-poor members of the greenschist facies or more commonly with quartz in the glaucophane schist facies (Fig. 2). Jadeite is formed by two common reactions:

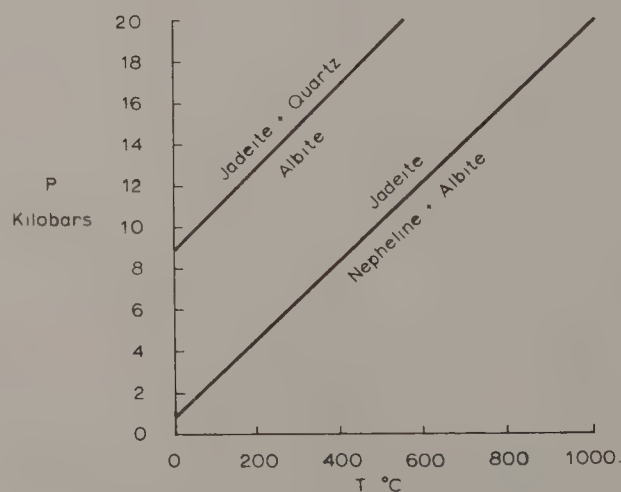
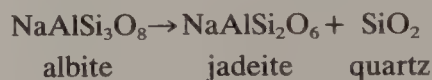


FIGURE 2. Regions of stability of jadeite.

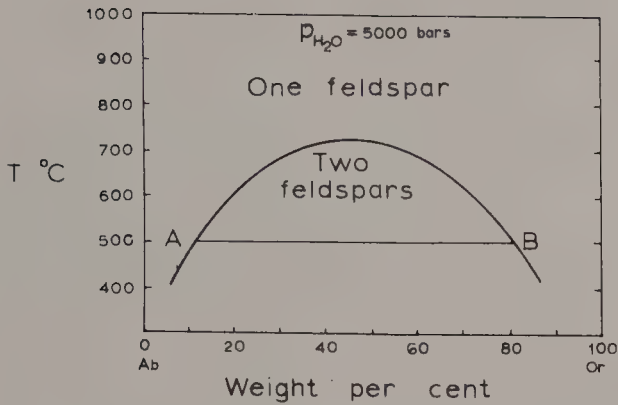
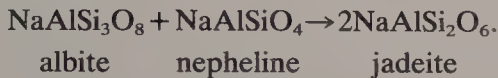


FIGURE 3. Unmixing of alkali feldspars; at 500°C a system along line A-B will consist of two feldspars of compositions A and B.

and



Jadeite is a dense phase and the above reactions represent a decrease in volume. Another example of the same type occurring between the sanidine facies and pyroxene hornfels facies is the transition from anorthite + wollastonite to grossular garnet with increasing pressure. In the vast majority of cases, the phase boundaries have a positive slope, as seen in Fig. 2.

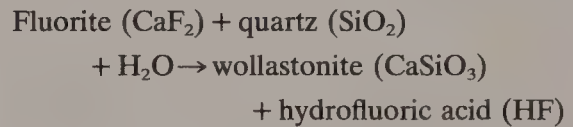
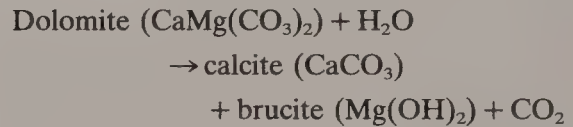
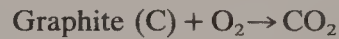
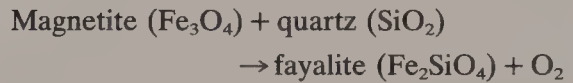
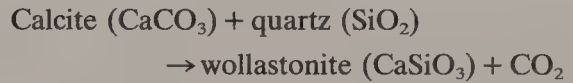
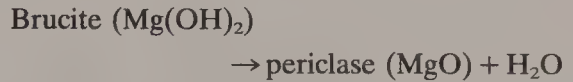
The conditions under which a given phase (e.g., jadeite) may form depends on the bulk chemistry of the system and the maximum field of stability will be obtained in a system whose composition is that of the pure phase. Thus, in using such phases as P - T markers, the reaction that produces or eliminates them must be known.

A second important class of solid-solid reactions involves mixing and unmixing processes. A simple case is provided by the alkali feldspars (Fig. 3). In some rocks of low temperature metamorphism, albite and K-feldspar may coexist. But in the granulite facies, the temperature is sufficiently high that total mixing occurs. Such mineral pairs thus show gradual approach in chemical composition with increasing temperature of metamorphism and may be sensitive markers of temperature. As the volume changes are often small, pressure may have only slight influence. Unmixing at low temperatures also occurs with the classic solid-solution series—the plagioclase feldspars—and two-plagioclase rocks may be quite common.

Solid-gas reactions

One of the most general types of metamorphic reactions involves the loss or addition of

gaseous species, most commonly H_2O and CO_2 . Many of the critical reactions involving the classification of metamorphic facies depend on such reactions. The following are some typical reactions.



The conditions under which such reactions may take place depend on total pressure, temperature, and the partial pressure of the gas involved. This leads to difficulty in interpretation, particularly of reactions in which two gases

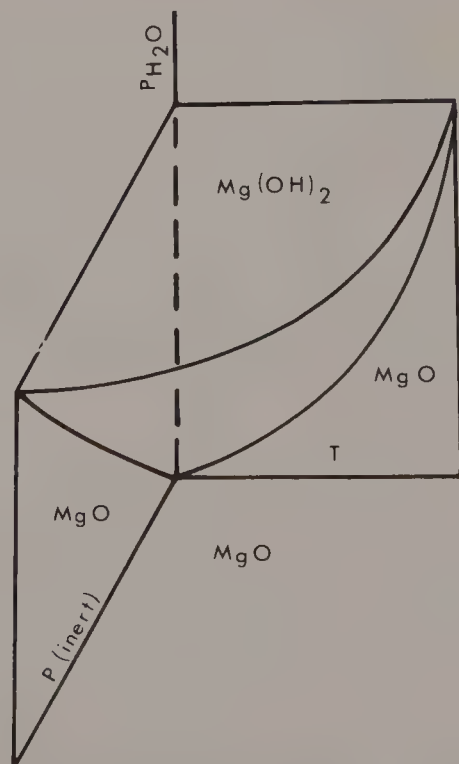


FIGURE 4. Phase diagram for the system $\text{MgO-H}_2\text{O}$ with variable T , $P_{\text{H}_2\text{O}}$ and P_{TOTAL} .

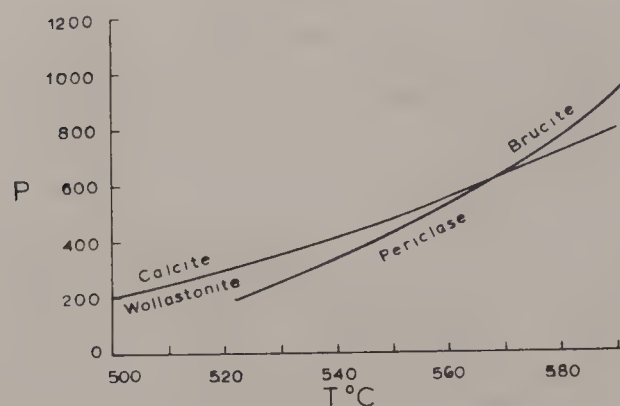


FIGURE 5. Vapor pressure curves for reactions $\text{CaCO}_3 + \text{SiO}_2 \rightarrow \text{CaSiO}_3 + \text{CO}_2$ ($P = P_{\text{CO}_2}$) and $\text{Mg(OH)}_2 \rightarrow \text{MgO} + \text{H}_2\text{O}$ ($P = P_{\text{H}_2\text{O}}$); P in bars.

are involved and the exact composition of the gas phase must be known. For a simple reaction, such as that involving brucite (Mg(OH)_2) and periclase (MgO), a complete phase diagram would appear as in Fig. 4.

Most metamorphic petrologists agree that the simplicity of patterns observed during dehydration reactions of noncarbonate sediments and loss of CO_2 from highly calcareous sediments suggests that partial pressures of the dominant gases and total pressures (load pressure) acting in the system are not completely independent. Thus, for a gas phase to escape from rocks with vanishingly small porosities at depth, P_{GAS} must approximate to P_{LOAD} . So simple vapor pressure curves of the minerals may be used to obtain information on limiting conditions. Most such curves have a form similar to that for brucite or calcite-wollastonite (Fig. 5). But in some systems, particularly where a low density hydrate may pass to a high density anhydrous phase, very high water pressures may lead to dehydration (Fig. 6).

Oxide vapor pressures are frequently very much smaller in magnitude (Fig. 7), and this implies that quantitative removal of oxygen is difficult at these low values. Thus, many such systems maintain these original oxidation states virtually without change during progressive metamorphism.

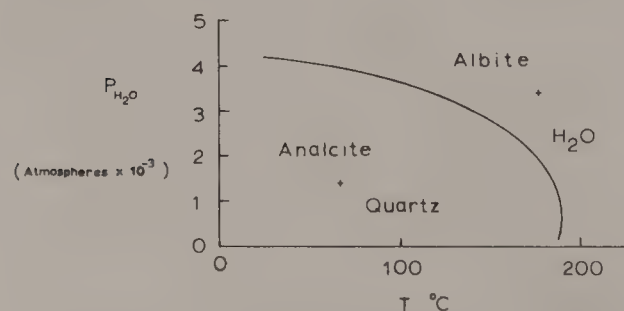


FIGURE 6. Possible phase diagram showing regions for the reaction analcime + quartz $\rightarrow$ albite + water.

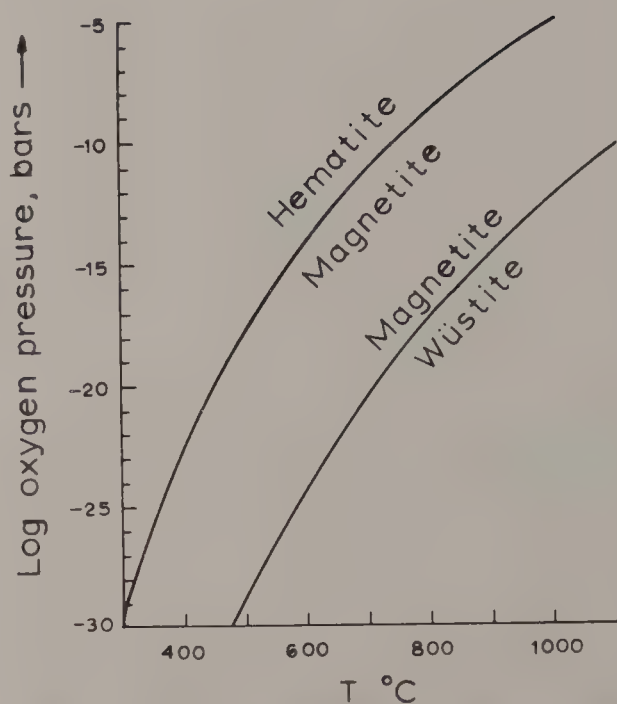


FIGURE 7. Oxygen pressures in equilibrium with pairs hematite-magnetite and magnetite-wüstite.

General statements

1. High P_{FLUID} or P_{LOAD} favor mineral assemblages of high density and systems containing hydrates and carbonates; coordination numbers in structures are large.
2. High temperature leads to low-density, volatile-free assemblages.
3. High temperature leads to maximum amounts of solid solution.

W. S. FYFE

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Cross-references: *Aureole*; *Burial metamorphism*; *Metamorphic mineral growth*; *Metamorphic petrology—use of thermodynamics*; *Reaction rates in metamorphism*; *Regional metamorphism*.

METAMORPHIC ROCK COMPOSITION—GRAPHICAL REPRESENTATION

Of the thirteen major chemical components of rocks, only six have very great effects on

mineral parageneses; these are SiO_2 , Al_2O_3 , CaO , $(\text{FeO} + \text{MgO})$, and K_2O . The other components either form small proportions of minor minerals (TiO , P_2O_5 , Fe_2O_3) or are considered to be mobile components open to the system (H_2O , CO_2). MnO proxies normally for FeO and Na_2O is most commonly present in plagioclase. In many metamorphic rocks, quartz is present so that SiO_2 can be ignored in these cases. Decrease in SiO_2 may lead to the formation of various SiO_2 -deficient minerals, but even these are rather limited in number. Thus, in rocks with excess SiO_2 the compositions may be plotted on two triangular diagrams suggested by Eskola (1915, see Miyashiro, 1973), the ACF and AKF diagrams. Other diagrams may illustrate more limited fields of composition.

Principles

The mineralogical phase rule deduced by V. M. Goldschmidt states that the number of minerals in stable coexistence is not more than the number of independent chemical components; thus, in a three-component system the number of stable minerals is either 1, 2, or 3 (Fig. 1). There may be a larger number of different minerals possible (six in Fig. 1) but a maximum of three are stable at any one time for a given composition. The differences between stability fields is denoted in such a diagram by the tie lines, e.g., at one temperature and pressure $\text{E} + \text{F}$ may be stable to give the stability fields shown in Fig. 1, while at a

different temperature and pressure $\text{D} + \text{C}$ may be stable rather than $\text{E} + \text{F}$, giving fields of $\text{D} + \text{F} + \text{C}$ and $\text{D} + \text{E} + \text{C}$, but not $\text{D} + \text{E} + \text{F}$ or $\text{F} + \text{E} + \text{C}$. A further complication arises due to the variation of composition of individual minerals with the temperature and pressure, and also within a facies, with the overall composition of the rock. Thus, the ideal composition of, say, clinopyroxene at granulite facies will not be a fixed end-number for that facies. Other methods, such as the AFM projection, are more suited to assessment of rock chemistry in such cases.

ACF diagram

The method of calculation of A, C, and F is to correct the molecular proportions (represented by square brackets) of $[\text{Al}_2\text{O}_3]$, $[\text{CaO}]$, and $[\text{FeO} + \text{MgO}]$ for the discrepancies caused by minor elements and accessory minerals; for example, apatite is assumed to contain all the P in the rock and will contain Ca in a fixed proportion to this P, so the $[\text{CaO}]$ is reduced by $3.33 \times [\text{P}_2\text{O}_5]$. Other constituents, such as MnO , are assumed always to be equivalent to one of the major oxides (in this case, FeO).

Eskola suggested subtracting proportions of FeO , Fe_2O_3 , and CaO based on the wt.% of the minerals ilmenite, magnetite, and sphene (determined by micrometric analysis). Further corrections are made after the oxides have been converted to molecular proportions to take into account the Al_2O_3 in alkali feldspars and the CaO in apatite and calcite. Winkler (1967) points out that corrections are also necessary for the $(\text{Fe}, \text{Mg})\text{O}$ in biotite and the Al_2O_3 in muscovite and paragonite (which is three times that present in orthoclase and albite, relative to the K_2O and Na_2O contents), if the ACF diagram is to be strictly accurate.

As polished sections are required for determination of the modal proportions of ilmenite and magnetite, and as the $\text{TiO}_2\%$ and $\text{FeO}/\text{Fe}_2\text{O}_3$ ratio of magnetite varies with metamorphic grade, there seems little point in treating TiO_2 and Fe_2O_3 any differently from P_2O_5 , Na_2O , and K_2O . Hence most workers now make the calculation on the basis of normative mineralogy. It is necessary to use appropriate normative minerals for the grade of rocks being investigated, so that TiO_2 should be assumed to be in sphene in greenschist facies rocks and in ilmenite in amphibolite facies rocks, but in eclogite and granulite facies the TiO_2 is normally present as rutile and can be ignored.

The procedure for calculating A, C, and F is as follows. Divide the wt.% of the oxides by their molecular weight, and for accurate

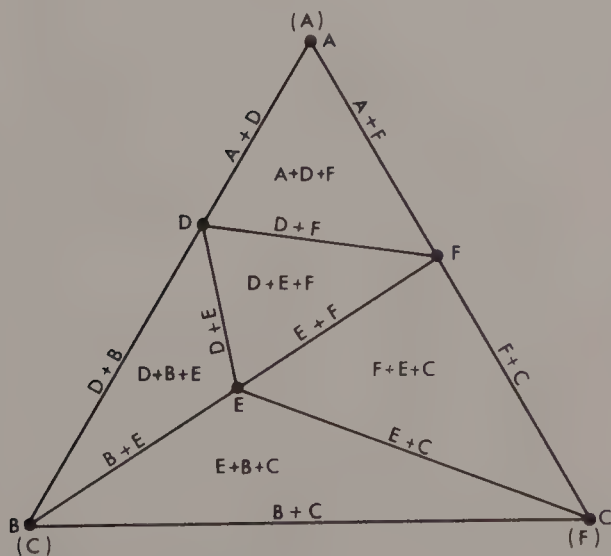


FIGURE 1. Composition-paragenesis diagram of a three-component system A-C-F in which minerals A, B, C, D, E and F are stable; possible assemblages are marked.

consideration of parageneses estimate the wt.% of biotite, muscovite and paragonite.

$$A = [Al_2O_3](-[Na_2O] - [K_2O] \text{ in plagioclase})(-3[K_2O] \text{ in muscovite})(+ [Fe_2O_3] \text{ if magnetite is not present})$$
$$C = [CaO] - 3.33[P_2O_5](-[TiO_2] \text{ if sphene is present})$$
$$F = [FeO] + [MgO] + [MnO](-[TiO_2] \text{ if ilmenite is present})(-[Fe_2O_3] \text{ if magnetite is present})(-3/8 \text{ biotite})$$

A, C, and F are expressed as percentages of A + C + F.

This calculation assumes that all Na₂O is in albite or paragonite and can therefore be ignored. It is thus unsuitable for plotting rocks of the glaucophane (lawsonite) schist and eclogite facies in which Na₂O plays an important role in amphiboles and pyroxenes.

Normally, silica-saturated compositions are shown (Fig. 2), but silica-undersaturated diagrams may be represented separately, although these are less accurate because parageneses may change with increasing undersaturation.

A point on the ACF diagram now indicates the minerals to be expected from that chemical composition at the particular metamorphic facies represented. For example, Goldschmidt found 10 assemblages in contact metamorphic pyroxene hornfels facies rocks at Oslo (Fig. 2), whereas Eskola found that some assemblages were the same but others different in regional metamorphic rocks at Orijärvi (Fig. 3); they thus belong to a different facies, the amphibolite facies. These diagrams show that only ferromagnesian assemblages are diagnostic between these two facies, the calcic and aluminous assemblages being identical.

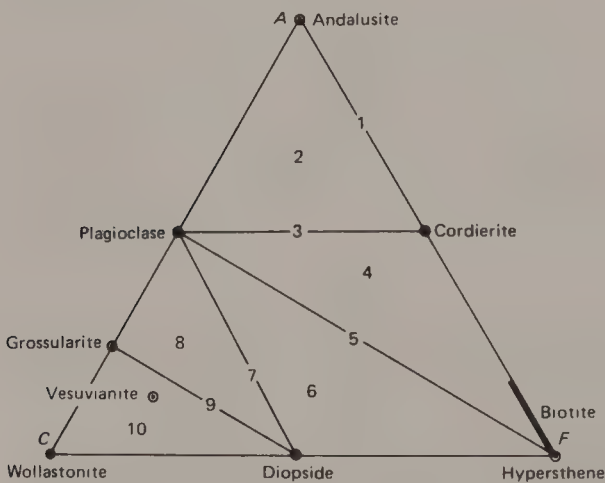


FIGURE 2. ACF diagram of pyroxene hornfels facies, Oslo, Norway, for rocks with excess SiO₂ and K₂O; quartz and orthoclase are possible additional components for all assemblages 1-10. (From Turner, 1968.)

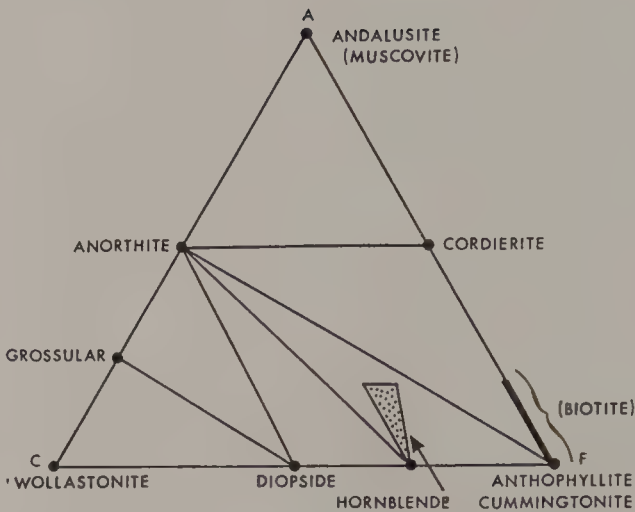


FIGURE 3. ACF diagram for amphibolite facies, Orijärvi, Finland, for rocks with excess SiO₂ (and K₂O); quartz is a possible additional component. (After Miyashiro, 1973.)

AKF diagram

The AKF diagram expresses the occurrence of minerals in potassic rocks. There are virtually no minerals containing both CaO and K₂O that cannot be represented by separate end-members, so it is quite feasible to plot two diagrams, ACF and AKF, to cover a very wide range of rock composition. The basic problems of computation are the same as for the ACF diagram, but those minerals containing Ca in addition to Al and Fe-Mg must also be corrected for.

Thus, A must ideally be corrected for grossular-andradite and zoisite, although it is normally only necessary to consider anorthite. F must ideally be corrected for such calcic minerals as diopside and hornblende, but the proportion of (Fe, Mg)O in such minerals is so variable that standardized corrections are probably not possible and both modal data and the chemistry of the individual pyroxenes or amphiboles should be known. It should be noted that both A and F are not the same as the A and F of the ACF diagram.

$$A = [Al_2O_3] - [Na_2O] - [K_2O] - ([Al_2O_3] \text{ in anorthite, zoisite, grossular, andradite}) + ([Fe_2O_3] \text{ if no magnetite present})$$
$$K = [K_2O]$$
$$F = [FeO] + [MgO] + [MnO] - [(Fe, Mg)O] \text{ in diopside and hornblende})$$

The standardized computation since Eskola's first usage has been to deduct the [Al₂O₃] equivalent to [K₂O] from [Al₂O₃], which results in microcline lying at the K corner of the triangle and the whole of the AKF diagram is

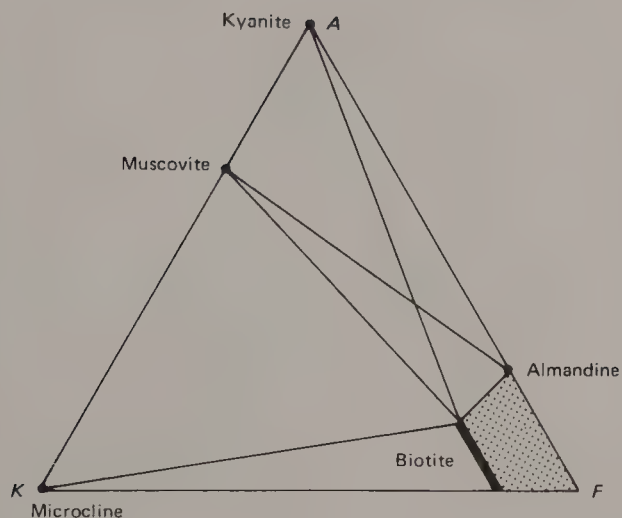


FIGURE 4. AKF diagram for kyanite zone, amphibolite facies, Glen Urquhart, Scotland, for rocks with excess SiO_2 ; quartz and plagioclase are additional phases in all assemblages; stippled field not found. (From Turner, 1968.)

diagrams; e.g., Fig. 4 represents the occurrence of the assemblages kyanite–almandine–muscovite–biotite, and kyanite–muscovite–biotite. (Francis, 1964, reference in Turner, 1981). Three explanations are possible for such assemblages: (1) the assemblages represent disequilibrium assemblages and are thus not suitable for plotting on AKF diagrams; (2) they represent a range of temperature and pressure conditions and thus again do not represent one facies and are unsuitable; or (3) they represent cases in which one component on the diagram is acting as two or more components in the minerals plotted, e.g., the “F component” is acting as the separate components $[\text{FeO}]$, $[\text{MgO}]$, and even $[\text{MnO}]$ —this is the only case for which crossing tie-lines should be used to express otherwise self-contradictory assemblages, but such assemblages are best expressed on the AFM projection.

AFM projection

Since many pelitic schists contain assemblages of four components in the AKF diagram (e.g., kyanite–almandine–muscovite–biotite at amphibolite facies and chlorite–muscovite–stilpnomelane–microcline and chloritoid–chlorite–pyrophyllite–muscovite at greenschist facies, it is obvious that the AKF diagram is not a suitable diagram for demonstrating the variation in mineralogy with chemistry of pelitic rocks. Thompson (1957—reference in Miyashiro, 1973) developed the AFM projection from a tetrahedron of the four molar com-

then used (Fig. 4). If this were not done, microcline would lie midway between K and A and the triangle K–Microcline–A would be devoid of any normal minerals.

Significance of crossed tie lines

In certain rocks more than three components are found (other than quartz and sodic plagioclase), or three component assemblages are found that should be mutually exclusive. These are indicated by crossing tie lines on the

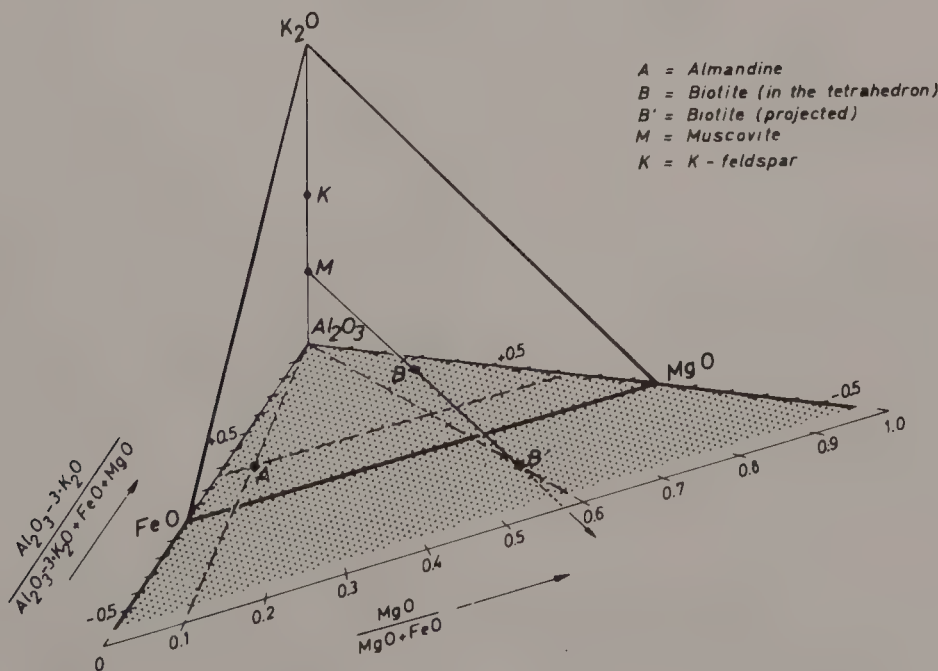


FIGURE 5. The tetrahedron $[\text{K}_2\text{O}]-[\text{Al}_2\text{O}_3]-[\text{FeO}]-[\text{MgO}]$ and the AFM triangle onto which the points within the tetrahedron are projected. (From Winkler, 1967.)

ponents $[Al_2O_3]$, $[K_2O]$, $[FeO]$, and $[MgO]$. In order to obtain a planar diagram, the compositions are projected on to the plane of $[Al_2O_3]$ – $[FeO]$ – $[MgO]$ (called the AFM plane), from the position of ideal muscovite ($[K_2O]:[Al_2O_3]=25:75$ —Fig. 5). This point is used because muscovite is stable over a wide P – T range and so most pelitic rocks contain muscovite as a principal constituent. In higher-grade rocks, where muscovite is absent, Barker (1961—reference in Miyashiro, 1973) has suggested that the K-feldspar point can be used as the origin of projection instead. The calculation of the position of any composition is made by calculating the ratios

$$\frac{([Al_2O_3] - 3[K_2O])/([Al_2O_3] - 3[K_2O] + [FeO] + [MgO])}{[MgO]/([MgO] + [FeO])}$$

and

$$\frac{[MgO]}{([MgO] + [FeO])}$$

The compositions of many minerals lie on the AFM plane and other common minerals are shown projected on to it in Fig. 6; its use in showing the stable minerals in a subfacies of the greenschist facies is shown in Fig. 7. This Thompson (AFM) diagram illustrates the principle of projection onto a plane of a tetrahedral system that has been used for more detailed consideration of varying rock composition. Its great advantage is that it allows the variation in mineral chemistry (particularly Fe/Mg ratio) with changing facies, to be accurately displayed. For nonpelitic (particularly basic) rocks, the Al_2O_3 – FeO – MgO – CaO

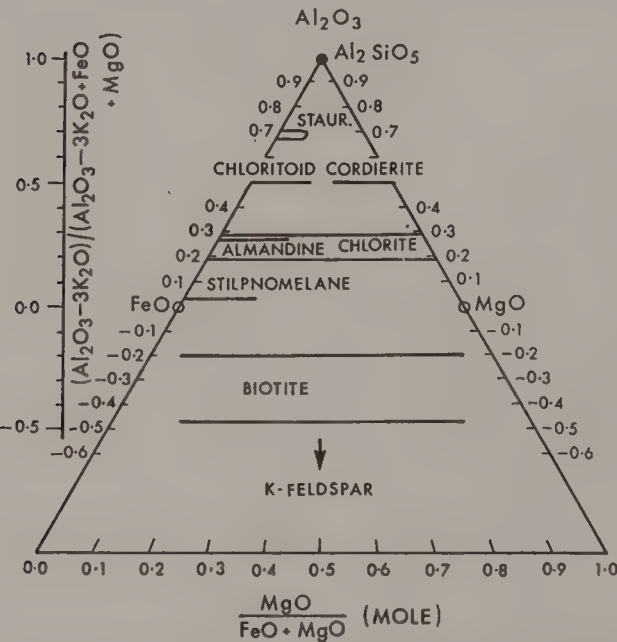


FIGURE 6. Thompson AFM diagram showing position of possible phases in equilibrium with quartz and muscovite. (After Miyashiro 1973.)

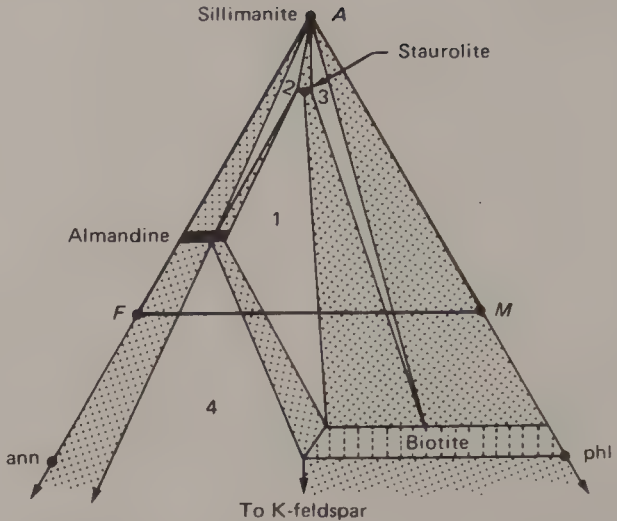


FIGURE 7. AFM diagram of the muscovite-sillimanite zone, New Hampshire; fields of two-phase assemblages stippled; three-phase fields blank; (1) almandine-staurolite-biotite; (2) almandine-staurolite-sillimanite; (3) sillimanite-staurolite-biotite; (4) almandine-biotite-orthoclase. (From Turner, 1968.)

tetrahedron has been used with anorthite (Robinson and Jaffe, 1969—reference in Spear et al., 1982) or epidote (Harte and Graham 1975; see also Spear et al., 1982) as the projection point (Fig. 8). It is also possible to project onto a plane parallel to the $[Al_2O_3]$ – $[CaO]$ edge from a point such as anorthite (Stout, 1972) to study the behavior in detail of the partitioning of Mg and Fe between the phases involved in the metamorphism of basic rocks (Fig. 9).

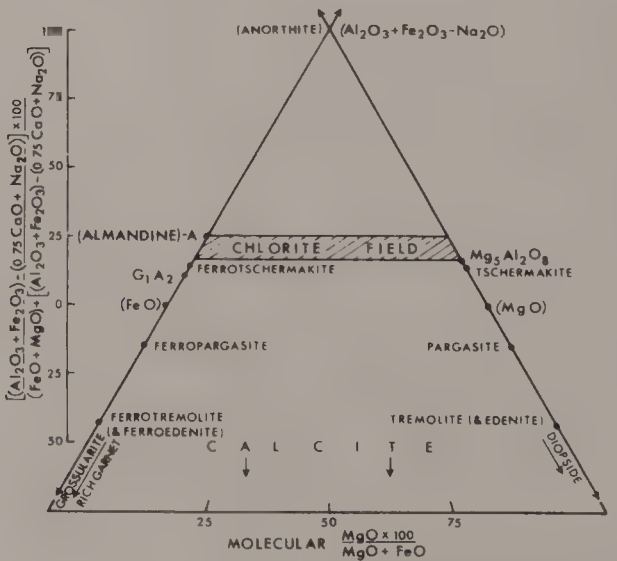


FIGURE 8. Projection from epidote onto AFM plane of $[Al_2O_3 + Fe_2O_3 - Na_2O]$ – $[CaO]$ – $[MgO]$ – $[FeO]$ tetrahedron with positions of major minerals. (From Harte and Graham, 1975.)

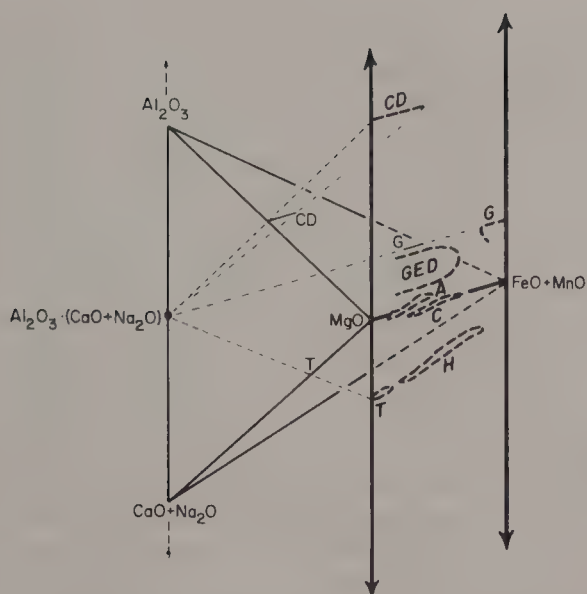


FIGURE 9. Projection of the system $[Al_2O_3]-[CaO + Na_2O]-[MgO]-[FeO + MnO]$ on to a plane parallel to the $[Al_2O_3]-[CaO + Na_2O]$ edge from plagioclase point; CD, cordierite; G, garnet; GED, gedrite; A, anthophyllite; C, cummingtonite; H, hornblende; T, tremolite. (From Stout, 1972.)

In all these cases, the detail given will only be valid if the data do not depart from the constraints imposed by the method. Extra components (e.g., K_2O or TiO_2 in the amphiboles or pyroxenes) will remove the plotted composition from the projection plane, and minerals rich in another component (e.g., K_2O or Na_2O) that contain a significant proportion of one of the end members must also be allowed for.

These and other projections are reviewed by Spear et al. (1982), who also give methods of dealing with more than four components by linear algebraic methods. Any number of components of a rock analysis can be fitted to whatever mineral end members are desired by the use of the equation of mass balance (Perry, 1967—reference in Spear et al., 1982), and the composition of the mineral end members may be changed to reflect the changing composition with facies. This development really makes ACF and AKF diagrams of little serious use in detailed metamorphic studies.

A. E. WRIGHT

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Cross-references: *Metamorphic facies*; *Metamorphism*; *Petrochemical calculations*; *Phase rule*.

METAMORPHIC ROCKS—OTHER TERMS

The main types of metamorphic rocks and their variants are dealt with in separate articles (see **Amphibolite**; **Buchite**; **Cataclastic rocks** (mylonite, phyllonite, pseudotachylite, etc.); **Charnockite**; **Eclogite**; **Gneiss**; **Granulite**; **Greenstone**; **Hornfels**; **Khondalite**; **Marble**; **Migmatite**; **Phyllite**; **Schist**; **Serpentinite**; **Slate**). The following are other metamorphic rock terms. Still others are covered by the use of the prefix meta, e.g., metadolerite, metapelite.

Arendalite—a garnetiferous rock.

Catawberite—mainly composed of talc and magnetite.

Collobrièreite—composed of grunerite, fayalite, garnet, magnetite, and some feldspar (Collobrieu, France).

Corundolite—see *Emery rock*.

Diabrochite—the product of metasomatic activity resulting from intensive penetration of ascending solutions or vapors (ichors) but without injection of visible granitic material as in migmatite.

Disthenite—almost entirely composed of kyanite (disthene) with some quartz; often associated with magnetite quartzite and amphibolite.

Ectinite—a metamorphic rock (particularly schist and gneiss) formed as the result of essentially isochemical regional metamorphism.

Emery rock—granular and mainly composed of corundum, magnetite, and spinel; formed by metamorphism of highly aluminous sediments (sanidinite facies) or by magmatic segregation.

Epidiorite—a metamorphosed gabbro or dolerite in which fibrous amphibole replaces pyroxene; generally massive but may be schistose; *epidiabase* is a (little-used) synonym.

Epidosite—consists of epidote and quartz, generally with other secondary minerals such as urallite and chlorite; commonly an altered igneous rock; epidote-quartz veins traversing

such a rock are also referred to as being epidosite.

Eulysite—fayalite is the most common mineral (the Mn-bearing variety in places); grunerite and almandine–spessartine also present; associated rocks are grunerite–garnet or hedenbergite–garnet assemblages; metamorphosed Fe (and Mn)-rich sediment often amongst gneisses (cf. *Peridotite*).

Fahlband—a metamorphic rock with a considerable proportion of Fe-ore or sulfides; a Scandinavian mining term.

Faserkiesel—quartz–fibrolite veins and knots.

Floittite—consists of biotite and other minerals typical of greenschist facies assemblages.

Garnetite—mainly composed of an aggregate of interlocking garnet crystals.

Gondite—composed of spessartine and quartz; a metamorphosed Mn-rich sediment (cf., eulysite) (Gonds, India).

Granofels—a medium- to coarse-grained granoblastic rock, generally with little or no foliation or lineation; used by some authors to designate a granulite-like rock not showing granulite facies mineralogy; *granolite* is also used.

Itabirite—a metamorphosed banded siliceous ironstone; original chert or jasper are represented by quartz that is separated by hematite–magnetite–martite layers (Itabira, Brazil).

Itacolumite—a micaceous sandstone or schistose quartzite that exhibits flexibility when split into thin slabs (Mt. Itacolumi, Brazil).

Jadeitite—composed of jadeite with small proportions of feldspar or feldspathoids; probably the product of high-pressure metamorphism of an alkali-rich igneous rock.

Kinzigitte—a granulite facies metapelite; garnet and biotite are essential constituents with varying proportions of quartz, K-feldspar, oligoclase, muscovite, cordierite, and sillimanite (Kinzig, W Germany).

Lavialite—a metabasite with relict labradorite phenocrysts in an amphibolitic groundmass; probably a metamorphosed basalt or basaltic tuff (Lavia, Finland).

Leptite—Fennoscandian term for a medium to fine-grained granular, quartzofeldspathic rock representing the high-grade metamorphic equivalent of acidic lava or tuff; cordierite-bearing leptites have been interpreted as having been derived from pelitic sediments; sometimes used as a synonym of quartzofeldspathic granulite; the term *halleflintgneiss* is a term formerly used in Sweden for a rock somewhat similar to leptite, but banded.

Leptynite—a term used by French geologists for feldspathic granulite like, or coarser-grained than, leptite; generally used now for neosome in granulite facies migmatite complexes, where it is

commonly spatially associated with charnockite; quartz, K-feldspar, biotite, and almandine are major minerals present.

Listwaenite (*liswänite*)—a schist with varying proportions of quartz, dolomite, magnesite, talc, and limonite; a much altered serpentinite.

Murasakite—a schist essentially composed of piedmontite and quartz (Murasko, Japan).

Para-amphibolite, *paragneiss*, etc.—amphibolite, gneiss, etc., derived from a sedimentary rock.

Ortho-amphibolite, *orthogneiss*, etc.—amphibolite, gneiss, etc., derived from an igneous rock.

Pinolite—magnesite (breunnerite) as crystals and as a granular aggregate in a matrix of phyllite or talc schist; the magnesite bodies resemble pine cones.

Quartzite—(1) granoblastic, consisting essentially of quartz, and representing metamorphosed sandstone; (2) sandstone or grit cemented by silica that has grown in optical continuity around the grains.

Soapstone—a rock essentially composed of talc with varying proportions of micas, chlorite, amphiboles, and pyroxenes and showing massive, schistose, or interlaced fibrous texture.

Sondalite—composed of cordierite, quartz, garnet, tourmaline, and kyanite.

Unakite—a metamorphosed hypersthene syenite with abundant epidote and subordinate orthoclase and quartz and accessory opaque oxides, apatite, and zircon (Unaka Range, Tennessee–N Carolina, U.S.A.).

D. R. BOWES

METAMORPHISM

Metamorphism is the mineralogical and structural adjustment of solid rocks to physical and chemical conditions (imposed at depths below the surface zones of weathering) that differ from the conditions under which the rocks in question originated (from the Greek meaning “change of form”). Generally metamorphism is isochemical (excluding water). Where the bulk chemical composition of the rock is changed, with addition and removal of material, the process is referred to as *metasomatism*. The branch of geology involving the study of metamorphic rocks is referred to as *metamorphic petrology*. This involves both description of the nature of metamorphic rocks (*petrography*) and determination of origin (*petrogenesis*).

On the basis of field occurrence metamorphic rocks are divided into three main categories:

contact metamorphic rocks, *regional* metamorphic rocks and *dynamic (cataclastic)* metamorphic rocks. Contact metamorphic rocks occur at or near the contacts of igneous intrusions with heat supplied to the country rock by the cooling intrusion being the most important agent causing metamorphic recrystallization: the process is often referred to as *thermal metamorphism*. Regional metamorphic rocks occur over large tracts of the Earth's crust and their recrystallization is associated with deformation that is often intense: recrystallization takes place as the result of *dynamothermal metamorphism*. Dynamic (cataclastic) metamorphism affects rocks in narrow zones in which particularly strong deformation has occurred, such as major fault and thrust zones. A special case is that of *shock metamorphism* near the impact sites of large meteorites.

Subcategories of metamorphism include *burial metamorphism* in which there is large scale recrystallization but no penetrative deformation, and *ocean-floor metamorphism* in which new mineral assemblages are produced as the result of static hydration particularly associated with burial and hydrothermal alteration at mid-ocean ridges.

D. R. BOWES

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Cross-references: *Burial metamorphism*; *Cataclastic rocks*; *Contact (thermal metamorphism)*; *Metamorphism—controls*; *Metamorphism—processes*; *Metamorphism—types*; *Metasomatism*; *Ocean-floor metamorphism*; *Regional metamorphism*; *Shock metamorphism*.

METAMORPHISM—CONTROLS

The various factors controlling regional or contact metamorphism are not different, it is only the emphasis that changes. Hence metamorphism can be considered as a continuum, with the various controls often acting together, but to different degrees or in different ways. For example, the effects of temperature are considered in the first section in an equilibrium context, while the role of temperature in a nonequilibrium or dynamic context is discussed in sections dealing with reaction kinetics and thermal cycle patterns.

Temperature

Temperature is one of the most important variables in metamorphism and is most clearly expressed in detailed isograd sequences in thermal aureoles. It is equally important in regional metamorphism where the variation in temperature across an area of prograde metamorphism is interpreted in terms of isograds marking the "incoming" of a mineral or mineral assemblages commonly formed by dehydration reactions that may be considered to be essentially isothermal. The sequence of minerals that form in regionally metamorphosed pelitic rocks, e.g., chlorite, biotite, garnet, etc. (in Barrovian zones), are thought to represent points on a series of univariant dehydration curves that cover the whole field of metamorphism. Isothermal curves in conjunction with near-isobaric univariant curves may be used to construct a petrogenetic grid covering the whole metamorphic P - T plane. Experimental and thermodynamic data on these reaction curves, when completely available, can then be used to give the exact temperatures and pressures for a given metamorphism. The fact that thermal and low-pressure metamorphism have nearly flat paths on P - T diagrams stresses the importance of the temperature control of the mineral assemblage.

Metamorphism is limited by two important temperatures boundaries, viz., at high temperatures, that of melting of rocks, and at low temperatures a rather indeterminate region above the temperatures producing diagenesis, i.e., above about 200°C. The high-temperature boundary will vary depending on the rock composition, water content, and, to a lesser extent, pressure, and probably lies between 650°C and 700°C for hydrous rocks. In anhydrous rocks, melting may not occur until much higher temperatures (1,000°C).

Pressure

Hydrostatic pressure (normally equated with depth) was in the early part of the 20th century

considered to be the principal factor determining the mineral assemblages and textures in metamorphic rocks (Grubenmann and Niggli, 1924, see Turner, 1981). More recently, the effect of hydrostatic pressure and temperature together has been accepted. Turner (1981), for instance, presents diagrammatically a schematic facies P - T grid that he uses to describe the petrologic and field aspects of metamorphism. Carmichael (1978, see Turner, 1981), in an attempt to formulate the importance of depth control, set up a scheme of bathozones based on critical assemblages lying on either side of a reaction curve (isobaric) through an invariant point, determined from the intersection of specific equilibria calibrated by thermodynamic or experimental data. The reactions (dehydration, hydration, polymorphic) relate to pelitic compositions and involve all three aluminosilicates. "Ideally, mapping of bathograds should provide a means of 'filtering out' the effects of temperature and bulk composition on metamorphic grade, so that the effect of pressure can be clearly perceived" (Carmichael, 1978, p. 793). The relationship of bathograds and isograds allows modeling of regional P - T regimes, specifically postmetamorphic folding and uplift or thermal doming, as well as determination of the mean geothermal gradient at the intersection of a bathograd and isograd. However, complications involving metastable persistence, experimental difficulties, overstepping, stressed systems, and kinetics tend to blur the detailed validity of the simple model, in spite of the evident gross control of an assemblage by pressure.

In natural systems, *nonhydrostatic stress* is the norm and has been invoked as an independent pressure variable, although today few petrologists would go so far as accepting the concept of stress and antistress minerals. However, some authors would suggest that nucleation may be greatly influenced by the state of strain and the velocity of deformation, so much so that nucleation and growth may not occur without deformation. More likely is the view that it might be restricted to assisting in the elimination of metastable or unstable phases, or merely accelerating reaction by increasing the active surface areas and the strain energy. Whatever the precise role of nonhydrostatic stress in nucleation and mineral growth, it is clearly important in small scale diffusion, where deformation may create marked free-energy gradients. The production of preferred orientation in metamorphic rocks is relatively clearly understood. Stress-controlled mechanisms include translational gliding, shape orientation in a flowing matrix, and thermodynamically stable crystallization (Flinn, 1965).

Overpressures, which assume that stress differences can be sustained at depth over the crystallization period, have often been proposed to account for the presence of high-pressure minerals in rocks. The invocation of overpressures is partly due to the inconsistencies in the experimental data, particularly of the aluminum silicate triple point, which has slowly but surely moved to lower pressures and temperatures in the last two decades, such that it may not now be necessary to invoke the concept. Although tectonic overpressures are certainly feasible, for short times at least, there seem to be no assemblages that unequivocally demand an explanation involving overpressures.

From the evidence of deep boreholes, *fluid pressure* (P_F) is frequently thought to approach the load pressure (P_L), and this equivalence is a basic assumption of a simple model of metamorphism (Turner, 1981). For instance, P_{H_2O} is expected to approach P_L , so that a series of dehydration reactions may readily be used to define the conditions of metamorphism. However, in some instances this may not be so and the position of a univariant dehydration curve relating to a given "isograd" will change with varying P_{H_2O} at a given P_L . If it is accepted that P_{H_2O} varies, then its value must be known in order to locate the position of the univariant curve on a P - T diagram. If CO_2 or other gas is added to the fluid system, it will have a similar effect on the equilibrium temperatures of the simple dehydration or decarbonation reaction, so that it is even possible that isograds may intersect (Carmichael, 1970, see Turner, 1981).

Not only will a specific gas phase determine the mineralogy in a complex manner in natural systems, it may also change with time. For example, in a thinly bedded limestone-shale sequence, the P_{H_2O}/P_{CO_2} ratio may be very large initially, then as decarbonation becomes dominant it will decrease to a minimum, when P_{CO_2} may be approximately equal to P_F . As decarbonation ceases, the ratio may change to a final value where $P_{H_2O} = P_F$.

Another important control of the metamorphic assemblage is the *oxygen fugacity*, which may well determine the assemblage that crystallizes. Almandine becomes unstable with increasing f_{O_2} , although it will tend to be stabilized in rocks by the addition of Mn and Ca. In extreme cases, for example, the distribution of isograds defined by the Fe-containing minerals may lose significance in relatively oxidized parageneses (Ganguly, 1968). Frequently, natural systems contain other species such as S, which also affect the assemblage, the effect on the silicate composition and amount being similar to that of oxygen. Implied in this discussion is that O_2 is a mobile

component, as suggested by the decrease in graphite in rocks with increasing grade. However, in some circumstances it is clearly buffered by the solid assemblage.

Although metamorphic petrologists tend to the opinion that $P_F \leq P_L$, mainly because experimental data relate to these conditions, studies of fine structure in metamorphic rocks suggest that P_F may be greater than or equal to the minimum principal compressive stress (Etheridge et al., 1983). Under such conditions, microfractures and grain-boundary bubbles and tubules may constitute a porosity. Deformation would produce interconnected porosity and marked permeability, which may be strata-bound or structurally constrained. Mass transfer and isotopic exchange indicates that fluids are essential elements in metamorphic reactions, mass flow, and deformation at medium and low grades of metamorphism. Etheridge et al. (1983) maintain that, as a result of the permeability and high P_F , convection may occur within capped layers. The more efficient transport of heat upwards by advection may have important consequences in the evolution of metamorphic terranes, parts of which are difficult to model by conduction alone, i.e., those characterized by andalusite-sillimanite assemblages.

Composition

Composition has a clear control on the metamorphic assemblages at any given temperature and pressure. Field geologists have often recognized that certain rock compositions are more "sensitive" to metamorphism than others. Thus, pure limestones and pure quartzites are not good indicators of metamorphic conditions, as the products are mineralogically similar to the original rocks. Basic rocks have been used with much success as well as ultrabasic, siliceous, calcitic, and dolomitic rocks, but to date it is still pelitic and graywacke compositions that have shown the greatest utility. The compositional control on minerals such as staurolite and chloritoid has long been recognized, although there has been some dispute over the exact constraints on the chemical composition (Hoschek, 1967, see Turner, 1981). Previously, the absence of minerals of this type has often been related to P - T conditions, particularly pressure; now it can be ascertained with more certainty whether the absence is due to an unfavorable composition.

Of considerable importance in metamorphic facies analysis are possible compositional constraints on the presence of the aluminum

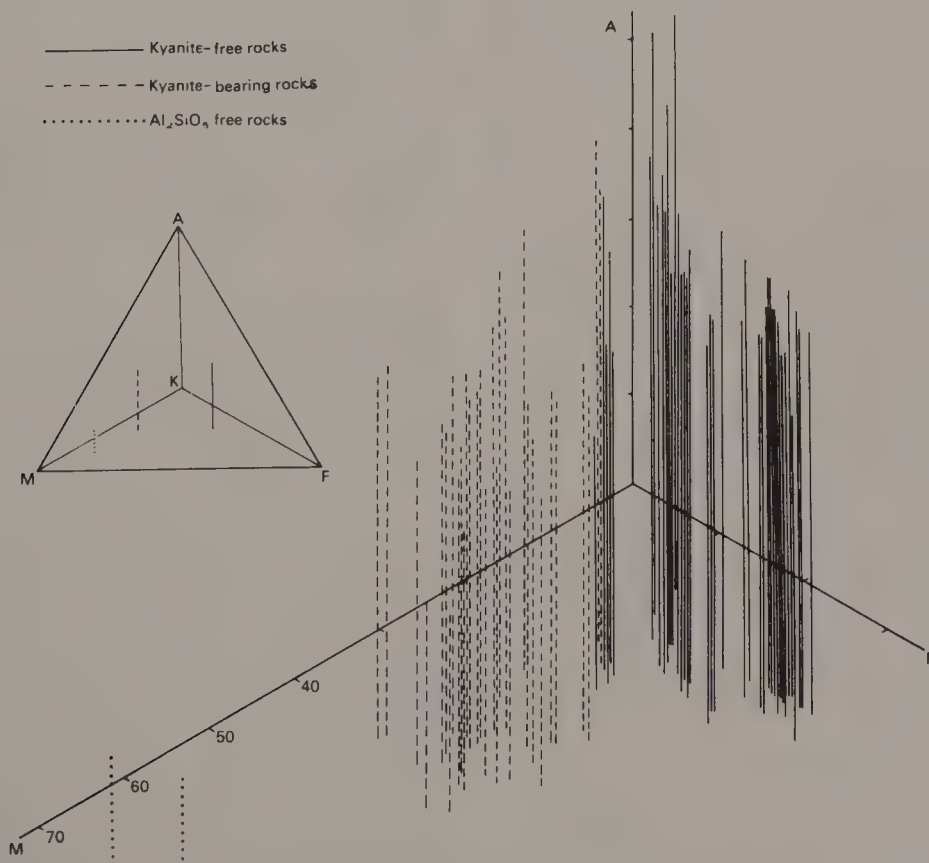


FIGURE 1. AKFM plot of rocks containing aluminum-silicate minerals from all aureoles showing the compositional restriction of the kyanite-bearing rocks.

silicates, particularly andalusite and kyanite. In Donegal, Eire, Naggar and Atherton (1970, see Turner, 1981) showed that kyanite is confined to rocks with a relatively high M/FM value (Fig. 1). The absence of kyanite in some rocks was interpreted in terms of a purely compositional constraint.

In order to minimize the compositional control, it would be necessary to study a full range of compositions in a thinly bedded sequence through varying grades. Such a situation is rarely realized in practice, for it is often found that stratigraphic elements in metamorphic areas are parallel or subparallel to isograds.

Reaction kinetics

Equilibrium is generally assumed in the use of isograds to establish variation in the crystallization temperatures and pressures in a given terrane. The general utility of AKF, ACF, etc., diagrams on a gross scale indicates this assumption is valid. However, it has been questioned by various authors, particularly Yoder (1952, see Turner, 1981), who thought that the appearance of an isograd could well be due to kinetics of growth rather than the attainment of equilibrium. This may well be true at the lowest grades, where reactants are highly unstable, e.g., basalts and clay-rich rocks. However, at higher grades, isograds, which often relate to dehydration reactions with high reaction entropies, are unlikely to be controlled by kinetics; furthermore, the long duration of a metamorphic cycle and textural evidence all suggest that kinetics have a minor influence on the position of an isograd.

There appears to be sufficient time during regional metamorphism to allow equilibrium to be reached, although it is known from microprobe work on garnets, etc., that homogeneous equilibrium is not common. Many porphyroblasts do not continuously react with the matrix during growth and in systems where depletion occurs zoning will result. By such processes it may be possible to move the reaction system into a new compositional field where minerals not in equilibrium with the earlier phases may later crystallize (Atherton and Brotherton, 1974). Some complex assemblages may be explained in this manner, while the metastable persistence of chemical gradients in minerals may be used to plot the P - T path of crystallization.

The extent of metastability in rocks is of concern because interpretations are based on experimental data on equilibrium assemblages. The problem is particularly acute with regard to the polymorphs of Al_2SiO_5 , for such solid-solid

transitions have very small entropy and free-energy changes, so that metastable phases may crystallize rather than the equilibrium phase. Metastable crystallization of andalusite in the kyanite field has been invoked in an occurrence in British Columbia, Canada. However, metastability should be more likely at low grades, particularly in zeolite facies rocks, yet it must be admitted that there is a singular consistency in mineralogy over large areas of pregreenschist metamorphosed areas, which suggests equilibrium was often approached, and that metastable crystallization is rare. Metastable persistence, however, undoubtedly occurs in many regional and thermal environments, particularly with regard to the aluminum silicate polymorphs and porphyroblast phases.

The occurrence of the three aluminum silicate polymorphs together in a rock was thought either to indicate crystallization at or very near to the triple point, or to be due to impurities extending the fields in which more than one phase could grow. Now it is clear that such an occurrence may also be due to metastable persistence of the lower temperature forms as the rock crystallized through an essentially isobaric path below the triple point (Naggar and Atherton, 1970, see Turner, 1981). Thus in the aureoles of granites in Donegal, Eire, kyanite is never seen to invert to andalusite or sillimanite on heating and sillimanite overgrowth on andalusite is common. Apart from metastability of the aluminum silicates, it is evident that metastability of other phases is common. Thus the survival of high-grade rocks is only explicable in terms of metastability associated with a lack of water, and the reaction kinetics of retrograde reactions.

Thermal cycle patterns

Many complex metamorphic sequences can be explained in terms of the dependence of the temperature field upon dynamic processes (Thompson, 1981). Thus, metamorphic crystallization and mineralogical sequences in space and in time are controlled by large-scale dynamic changes such as crustal thickening by overthrusting, continental collision, magmatic accretion within or at the base of the crust, and crustal thinning. Perturbed geotherms may be evaluated for these dynamic situations, and also their evolution with time as erosion and upward transport bring the material to the surface. P - T -time paths for individual samples within the pile will show unique paths back to the surface. Important parameters affecting the heating of modified crust include thermal conductivity, heat sources or sinks in various crustal layers, heat flow from beneath the crust,

depth of thickening, duration of uplift, and elapsed time after thrusting, etc., until erosion begins (England and Richardson, 1977). Smaller-scale perturbations of the geothermal gradient will occur against high-temperature basic plugs, for example, and may be characterized by extreme gradients. In these cases, time at maximum temperature, heat content of the hot bodies, and temperature contrast, will be important determining parameters, and overstepping of reactions may well be the norm.

M. P. ATHERTON

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Cross-references: *Aureole*; *Contact (thermal) metamorphism*; *Nucleation of metamorphic minerals*; *Reaction rates in metamorphism*; *Regional metamorphism*; *Retrograde metamorphism*.

METAMORPHISM—PROCESSES

The following terms have been used. Assessment of metamorphic processes is set out in Vernon (1976).

Anatexis—see **Anatexis**.

Annealing—toughening by recrystallization.

Anogenic—relating to plutonic metamorphism or replacement.

Blastic deformation—a process operative in dynamothermal metamorphism; recrystallization according to Riecke's principle results in elongation of previously existing minerals, and the growth of new minerals, in the plane normal to the direction of greatest pressure.

Charnockitization—see **Charnockite**.

Crystalloblastesis—deformation accomplished by metamorphic recrystallization.

Degradation recrystallization—recrystallization resulting in a decrease in crystal size.

Dissociation—the breakup of a compound to form two or more other substances (e.g., CaCO_3 becoming $\text{CaO} + \text{CO}_2$); the temperature of breakup is referred to as the dissociation point.

Epigenesis—the change in the mineralogical character of a rock as the result of external influences operating near the surface of the Earth, e.g., mineral replacement during zeolite facies metamorphism.

Granulation—formation into grains or granules, particularly as the result of tectonism under conditions such that no visible openings result.

Granulitization—reduction of components of a rock to grains, commonly as the result of ductile deformation; an extreme result is the development of mylonite (see **Cataclastic rocks**).

Heterogeneous pressure—see *Riecke's principle*.

Mesitism—transformation of chemically different rocks under the same P – T conditions leading towards homogenization.

Metablastesis—recrystallization and growth of a mineral, or group of minerals, associated with essentially isochemical metamorphism.

Metamorphic diffusion—migration, by diffusion, of materials from one part of a rock mass to another during recrystallization.

Metamorphic segregation—concentration of a particular mineral, or groups of minerals as the result of chemical and physical conditions operative during metamorphism, e.g., “blind” quartz veins in schists and compositional banding in gneisses.

Metasomatism—see **Contact metasomatism**; **Metasomatism**; **Pyrometasomatism**.

Migmatization (feldspathization, granitization)—see **Migmatite**.

Mimetic crystallization—the reproduction of a pre-existing structure (e.g., bedding) by metamorphic recrystallization.

Mylonitization—see **Cataclastic rocks**.

Petroblastesis—formation of rocks mainly as the result of crystallization of diffusing ions.

Recrystallization—formation of new, crystalline material as the result of metamorphic processes.

Riecke's principle—the principle that solution of a mineral tends to occur most readily at points where external pressure is greatest and crystallization occurs most readily where external pressure is least; used to explain changes in mineral shapes during metamorphic recrystallization.

zation, e.g., the development of augen. (The principle is named after E. Riecke, 1845–1915, although it was set out by H. C. Sorby in 1863.)

Strain recrystallization—recrystallization in which a deformed mineral changes to a mosaic of undeformed crystals of the same minerals.

D. R. BOWES

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METAMORPHISM—TYPES

Accidental—metamorphism adjacent to igneous masses (see **Contact (thermal) metamorphism**).

Additive—see *Allochemical*.

Aggrading neomorphism—metamorphism in which crystal size increases.

Allochemical—metamorphism accompanied by addition and removal of material (see **Metasomatism**); the bulk chemical composition of the rock is changed.

Anamorphism—prograde metamorphism at considerable depth in the crust (in the anamorphic zone); relatively dense minerals form at the expense of ones of lower density and rock flowage takes place.

Anchimetamorphism—changes in the mineral composition of rocks under the temperature and pressure conditions prevailing between the Earth's surface and the zone of true metamorphism (in zones of groundwater circulation and weathering).

Autometamorphism—see **Alteration, replacement**.

Burial—see **Burial metamorphism**.

Cataclastic—see **Cataclastic rocks**.

Coalesce—a type of aggrading metamorphism in which small crystals are converted to larger ones by gradual enlargement while maintaining a uniform crystal size.

Contact—see **Contact (thermal) metamorphism**.

Degrading neomorphism—metamorphism in which crystal size decreases.

Depth—see **Burial metamorphism**.

Diastathermal—metamorphism associated with extensional tectonism in regions of relatively high heat-flow, e.g., continental rifts.

Dislocation—see **Cataclastic rocks**.

Dynamic—see **Cataclastic rocks**.

Dynamothermal—see **Regional metamorphism**.

Eometamorphism—early or initial metamorphism.

Geothermal—static recrystallization associated with the deep burial of overlying rocks and a regular downward increase in temperature (see **Burial metamorphism**).

Hydrometamorphism—rock alteration, not at high temperatures and pressures, by material added or removed by hydrous solutions.

Injection—metamorphism accompanied by intimate injection of magma (usually granitic) in zones near the contacts of deep-seated intrusions.

Isochemical—metamorphism with no change in bulk chemical composition.

Kinetic—rock deformation without chemical reconstruction or recrystallization to form new minerals (see *Dynamic*).

Load—see **Burial metamorphism**.

Local—contact metamorphism adjacent to a heat source (e.g., igneous body), as distinct from regional metamorphism; rarely used in relation to cataclastic metamorphism in a narrow zone.

Low-rank—recrystallization under conditions of low to moderate temperatures and pressures.

Neomorphism—all transformations between one mineral and itself or a polymorph, whether the new crystals are larger (aggrading) or smaller (degrading), or simply differ in shape from previous ones, or represent a new mineral species; the term includes processes of inversion, recrystallization, and strain recrystallization, in which the bulk composition remains essentially constant.

Ocean-floor—see **Ocean-floor metamorphism**.

Orogenic—a type of regional metamorphism, generally low grade, resulting from regional heating and deformation connected with orogenic movements.

Plutonic—deep-seated regional metamorphism at high temperatures and pressures, often accompanied by strong deformation and with partial fusion—*anatexis* and development of injection phenomena, as in migmatite complexes.

Pneumatolytic—see **Contact metasomatism; Pneumatolysis**.

Porphyroid neomorphism—a type of aggrading neomorphism in which small crystals are converted to large ones by growth of a few large crystals in and replacing a static matrix, as in the replacement of an aragonite shell by a calcite mosaic.

Prograde—associated with increasing temperature and metamorphism.

Pyrometamorphism—see **Pyrometamorphism**.

Regional—see **Regional metamorphism**.

Retrograde—see **Retrograde metamorphism**.

Shock—see **Shock metamorphism**.

Static—a type of regional metamorphism resulting from the action of heat and fluids at high lithostatic pressure that results from load and not orogenic deformation (see **Burial metamorphism**).

Thermal—mainly associated with igneous intrusions (see **Aureole**; **Contact (thermal) metamorphism**; **Pyrometamorphism**) but also with regional environments with high thermal gradients (see **Metamorphic facies series**).

Ultrametamorphism—processes operative at high temperatures and pressures at which partial fusion takes place, leading to the formation of migmatites (see **Anatexis**; **Migmatite**).

D. R. BOWES

METASOMATISM

Metasomatism is the process of altering the composition of a rock, either by the addition or subtraction of chemical elements. The term is usually reserved for subsurface processes and, as commonly used, excludes weathering and diagenetic processes. According to Lindgren (1928) it is the practically simultaneous capillary solution and deposition by which a new mineral of partly or wholly different chemical composition may grow in the body of an old mineral or mineral aggregate.

The migration of chemical components within the Earth's crust requires potential gradients in the partial molar free energies (chemical potentials) of the moving components. Origins of such gradients may be thermal gradients, stress gradients, or variations in bulk compositions leading to variations in the compositions of single phases (minerals and fluids).

Proposed mechanisms for such migrations have been by means of fluid media, or by diffusion in the solid state. The ubiquity of voids and fracture systems in most rocks, coupled with faster rates of diffusion and flow in fluids, has led to almost universal acceptance of the fluid phase as the dominant metasomatic agent. An additional constraint on the mechanism may be the maintenance of constant volume, an interpretation based on relict textures found within the altered rock.

Korzhinskii (1957) has formulated a variation of the mineralogical phase rule for systems with perfectly "mobile" components. This treatment distinguishes those components (c) whose chemical potentials are defined by the bulk composition of the system at the temperature and pressure in question, as opposed to those components (m) whose chemical potentials are defined by a "reservoir" outside the system under examination. The number of phases

(minerals) (ϕ) expected if such system approaches "local equilibrium" is

$$\phi = c - m.$$

Application of this rule to the monominerallic and biminerallic zones characteristic of many metamorphic assemblages and sulfide deposits has led to reasonable estimates of the mobile components involved in the alteration that produced the presently observed assemblage.

Another test, where possible, of metasomatism by a pervasive fluid phase, is a comparison of the chemical potentials of the constituents of the mineral assemblages. Wide variations in the relative activities of components, such as H_2O , CO_2 , O_2 , H_2 , C, or S_2 , tend to discount the presence of a pervasive fluid phase causing large changes in the composition of the host rocks.

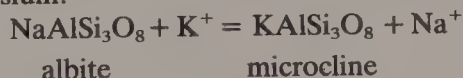
Geologic phenomena which have been successfully explained in terms of metasomatic processes are skarns, serpentinites and their contact phenomena, anomalous compositions in metamorphic terranes, contact alteration at the margins of intrusive igneous rocks, and the alterations associated with vein deposits. Certain variations in sedimentary rocks which occur because of reactions with pore fluids have been called metasomatic, i.e., dolomitization, but these are more properly termed diagenetic.

Mineral assemblages resulting from metasomatic activity usually contain only two or three phases and have a typically zonal distribution, mono- or biminerallic zones being typical, and, when interpreted in terms of the Korzhinskii phase rule, such convergent zones are considered evidence of metasomatic activity.

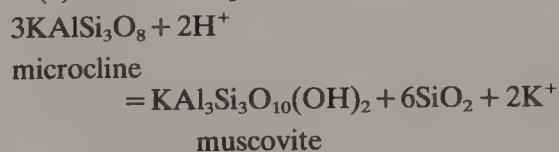
Alkali metasomatism is dependent on the alkali concentration and the ratio Na^+/K^+ within the metasomatizing fluids. Alteration haloes of the zoned assemblages

- (a) quartz-microcline-albite
- (b) quartz-microcline
- (c) quartz-muscovite
- (d) quartz-andalusite (pyrophyllite)

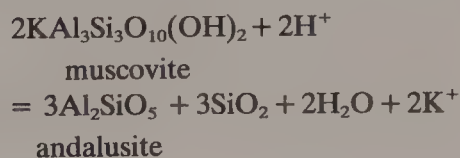
are examples of (1) replacement of sodium by potassium:



and (2) removal of potassium:



and



The mineral assemblages may be described by the system K_2O – Na_2O – Al_2O_3 – SiO_2 – H_2O , and, using the Korzhinskii phase rule, it can be deduced that in zone (a) Na_2O , K_2O , Al_2O_3 , and SiO_2 are inert, and in zones (b)–(d) Al_2O_3 and SiO_2 are the inert components.

As the ratios K^+/H^+ and Na^+/K^+ which determine the mineral assemblage in equilibrium with the fluid are very dependent on temperature, slight temperature gradients provide sufficient chemical potential gradients for metasomatism to take place.

The origin of some granitic rocks has been attributed to metasomatic processes. The arguments for such origins are based chiefly on the problem of providing “space,” the layered structure of gneisses, and the presence of relict structures conformable with structures in the enclosing host rocks. The heterogeneous nature of gneissic and batholithic mineral assemblages and the activities of components within such rocks are strong arguments against pervasive metasomatic activity. The similarity of silicate minimum-melting compositions with the compositions of batholiths and gneisses provides a more rigorous argument for the magmatic origin of the majority of granitic rocks.

Metasomatism takes place in some rocks adjacent to igneous intrusions (see **Contact (thermal) metamorphism**; **Skarn**). It may also affect extensive areas (regional metasomatism), with the introduction of fluids possibly related to partial fusion at depth. *Autometasomatism* relates to the alteration of an igneous rock mass by its own late H_2O -rich liquid fraction trapped within the rock, generally by an impermeable chilled border (see **Alteration, replacement**). A general term for a rock formed as the result of metasomatism is a *metasomatite*.

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Cross-references: *Contact metasomatism*; *Contact (thermal) metamorphism*; *Hydrothermal processes*; *Phase rule*; *Pneumatolysis*; *Pyrometasomatism*; *Skarn*.

METEORITES—PETROLOGY

Meteorites include stony, iron, and stony–iron types, most of which may be distinguished from terrestrial rocks by the occurrence of varying proportions of Fe–Ni metal and troilite (FeS). Over 90% of meteorites seen to fall are stony. A small proportion of stony meteorites together with all the irons and stony–irons are differentiates or mixtures of differentiated components produced by crystallization from a melt in a planetary or asteroidal interior. The remainder, the chondrites, have not undergone melting on a planetary or asteroidal scale since their formation ca. 4.5 Ga ago.

Classification of meteorites commonly incorporates the first letter of the place of discovery. It also may reflect the collection and cataloging practice of different museums and so varies from place to place. Suggested standard classifications, such as that of Dodd (1981) for chondrite meteorites (Fig. 1), must be used bearing these factors in mind.

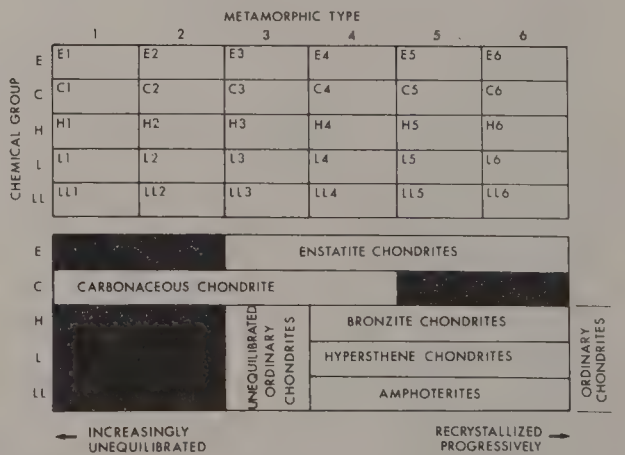


FIGURE 1. Classification of chondritic meteorites. (After Dodd, 1981.)

Chondrites

Chondrites are stony meteorites (*aerolites*) that almost invariably contain millimeter-sized

subspherical masses of silicate (more rarely metal or sulfide) known as *chondrules*. Formation probably occurred on the surfaces of small planetary bodies by the aggregation of chondrules and chondrule fragments, clasts of preexisting rocks, still-molten droplets, metallic grains, and fine dust. Most chondrites crystallized either during cooling from a high ambient temperature at the time of accretion or in a prograde thermal metamorphic event (Dodd, 1981); the end product is a rock with a texture approaching allotriomorphic granular in which the original quench textures have largely been erased (petrological type 7). The least-affected examples are composed of large numbers of chondrules with sharp outlines against fine-grained matrix (petrological types 1, 2, 3). In chondrules, rapid solidification is attested by the presence of isotropic glass containing skeletal crystals of olivine, or phenocrysts of low-Ca clinopyroxene showing polysynthetic twinning (Fig. 2a). Chondrules such as these typify petrological type 3 in the van Schmus and Wood classification (Fig. 1; Dodd, 1981). With increasing crystallinity, chondrule outlines tend to disappear owing to intergrowth, reaction with, and equilibration with the matrix. Stones in which chondrules are rare, marked only by indistinct outlines, are classed as petrological type 6. In these stones, primary monoclinic low-Ca pyroxene has all inverted to orthopyroxene, and crystalline oligoclase is present (Fig. 2b). Rarely, the process has completely erased chondrules, signaling the attainment of type 7. Petrological types 1 and 2 are reserved for low-temperature, mainly volatile-element-bearing or volatile-compound-bearing mineral assemblages; types 1 and 7 are therefore chondrule-free.

Superimposed on the petrological classification is a chemical one. Three major chemical groupings are recognized, viz. enstatite chondrites, ordinary chondrites (subdivided) and carbonaceous chondrites.

1. *Enstatite chondrites* (*E chondrites*) are highly reduced, having almost no Fe in silicates, and carry an unusual suite of carbide, nitride, silicide and sulfide minerals (Dodd, 1981); they have a low Mg/Si atomic ratio of about 0.8 and comprise two subgroups, with low and high total Fe (EL and EH groups), each apparently with its own "metamorphic" series, EL5–6 and EH3–5 (Sears et al., 1984).

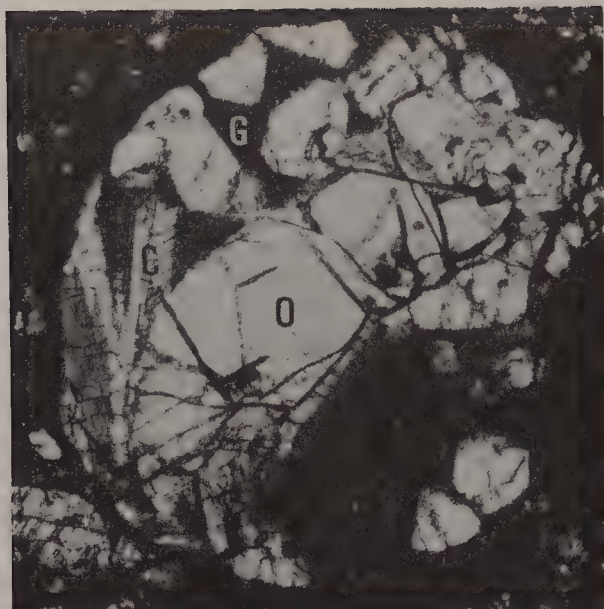
2. *Ordinary chondrites* are less reduced than the E chondrites and contain Fe-bearing olivine and pyroxene. The distribution of Fe between metal, silicate and sulfide results in a decrease in metal content compared with the E chondrites. The ordinary chondrites are further subdivided on the basis of total Fe content and oxidized Fe

content, reflected in different abundances of metal and in different olivine compositions.

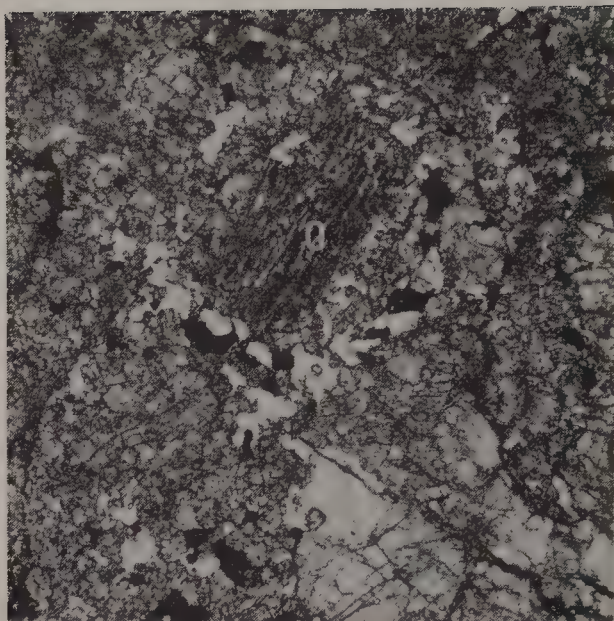
- 2a. *H-group* or *bronzite chondrites*, with about 28 wt.% total Fe, 20% Fe–Ni metal, and olivine Fa_{16-20} . "H" is for high total Fe, and H3–H6 are known.
- 2b. *L-group* (low total Fe) or *hypersthene chondrites*, with about 22% total Fe and 10–5% Fe–Ni metal and olivine Fa_{22-26} , are represented by L3–L7.
- 2c. *LL-group* (low total Fe, low metal) or *amphoterites*, with about 20% total Fe, <5% Fe–Ni metal, and olivine Fa_{27-31} , comprise LL3–LL6; most are breccias. Metal content generally decreases from H to L to LL and is to some extent inversely related to the abundance of olivine and to its FeO content. Decreasing metal content is accompanied by an increase in the proportion of Ni in the metal; taenite becomes more abundant relative to kamacite from H to L to LL.

3. *Carbonaceous chondrites* are the least reduced meteorites and have the highest Mg/Si atomic ratio of about 1.05. They are polymict breccias with high-temperature and low-temperature components, the latter resulting from in situ hydrothermal alteration of a petrological type 3 primary assemblage (Bunch and Chang, 1980). Chemically they may be subdivided into four types.

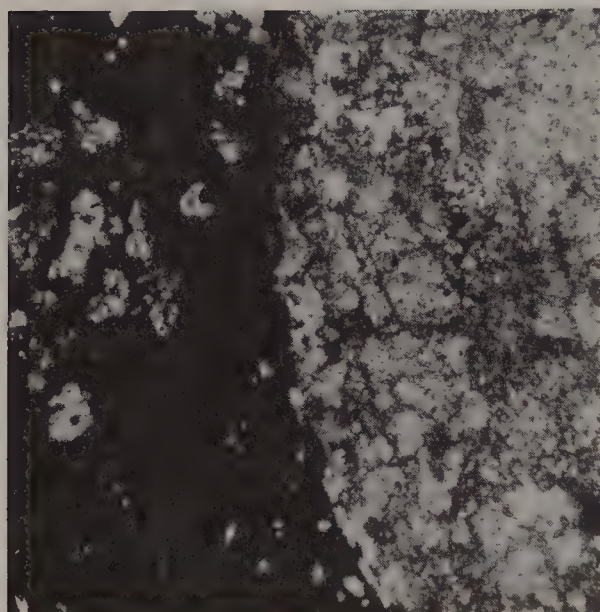
- 3a. *CI* ("I" for Ivuna) *members* lack chondrules and are richest in volatiles. The five known meteorites are all of petrological type 1 and have a small high-temperature component of euhedral olivine and magnetite crystals set in a groundmass composed of nearly amorphous hydrated silicate, plus calcite, gypsum, epsomite, and organic compounds: the organic compounds are now regarded as abiotic.
- 3b. *CM2* ("M" for Mighei) *representatives* have a groundmass like CIs, but additionally contain chondrules, rare inclusions rich in Ca, Al, and Ti, and a small metal component.
- 3c. *CV chondrites* ("V" for Vigarano) include petrological types 2–5 and are typified by the occurrence of large chondrules containing abundant opaque minerals together with inclusions rich in the refractory elements Ca, Al, and Ti (Kornacki and Wood, 1984). CVs have the highest Ca/Si and Al/Si ratios of all chondrite groups (Fig. 2c).
- 3d. *CO chondrites* ("O" for Ornans) are a rare group with small chondrules and rare Ca-, Al-, and Ti-rich inclusions; all are of



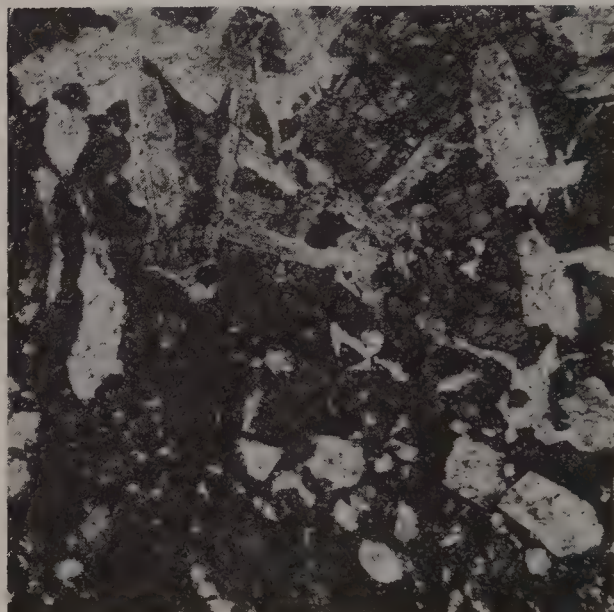
a



b



c



d



e



f

petrological type 3. Their nonvolatile-element abundances are like those of the CMs (Kallemeyn and Wasson, 1981).

Achondrites

The achondrites include metal-poor igneous rocks and regoliths that are aggregates of impact-produced debris of the primary igneous material. In the past, achondrites were subdivided into Ca-rich and Ca-poor groups, but this has no genetic significance and today various associations or clans are recognized, some of which suggest links with stony-irons and irons.

1. *Eucrites, diogenites, howardites (and mesosiderites)*. The eucrites are "basalts" composed of anorthitic plagioclase plus Fe-rich pigeonite and/or augite. Most are polymict breccias (Fig. 2d), but a few monomict eucrites and cumulates of plagioclase and pyroxene are known. Cumulates of a more magnesian low-Ca pyroxene, almost all of which are brecciated, are known as diogenites. A single dunitic meteorite (Brachina) completes the primary association. The howardites represent a regolith derived from the primary source rocks with the addition of an exotic chondritic component.

2. *SNC meteorites* comprise four *shergottites*, three *nakhlites*, and the unique olivine cumulate, *Chassigny*. Shergottites are shocked plagioclase-augite cumulates, the plagioclase

being andesine to labradorite, but mostly converted to the amorphous glass. The nakhlites are augite-olivine cumulates, with interstitial andesine. They are unshocked, and Nakhla has a crystallization age of about 1.2 Ga (Dodd, 1981). Chemical and isotopic evidence suggests a martian origin (Smith et al., 1984).

3. Three meteoritic stones found in Antarctica have textures and mineralogy consistent with lunar highlands breccias (see, e.g., Kurat and Brandstätter, 1983).

4. The *ureilites* are a group of about 10 olivine-pigeonite cumulates with intercumulus graphite. Shock has converted some of the graphite to diamond and reduction by carbon has produced Fe metal by reaction with the rims of the olivines. Ureilites are probably related to the CO chondrites, but the origin of the graphite is disputed; it could be primary or could have been introduced by the projectile that caused the shock (Begemann and Ott, 1983 and references therein).

5. *Aubrites*, or *enstatite achondrites*, are coarse-grained meteorites all but one of which are brecciated, and some contain implanted solar wind, testifying to residence on a planetary surface. Minor constituents include oldhamite (CaS) indicative of a reduced state like that of the E chondrites (Dodd, 1981).

6. *Angra dos Reis* (stone-*angrite*) is unique, a cumulate of aluminous, Fe-rich, fassaitic pyroxene plus olivine with 1.3 wt.% CaO. It has a crystallization age of >4.5 Ga (Dodd, 1981).

FIGURE 2.

(a) Chondrule in the H3 chondrite Brownfield (1937), typical of the petrological type 3, with olivine (O) and clinobronzite (C) phenocrysts in a glassy mesostasis (G); XN, $\times 36$.

(b) Appley Bridge LL6 chondrite; well-crystallized, petrological type 6, with a barred olivine chondrule (O) of which the rim is rich in clear, crystalline feldspar; note the overall granular texture and absence of well-defined chondrules; PPL, $\times 29$.

(c) Vigarano CV3 carbonaceous chondrite; on the right is an aggregate of pyroxene and melilite ($\text{Ca}_2\text{Al}_2\text{SiO}_7$) with some anorthite; the remainder consists of smaller chondrules and fragments of olivine and pyroxene crystals set up in a dark matrix; PPL, $\times 46$.

(d) Stannern eucrite, the upper half being a basaltic clast set in breccia which dominates the lower left; PPL, $\times 29$.

(e) The Rowton, group IIIA, medium octahedrite, iron meteorite; polished and etched, exhibiting a superficial, narrow, heat-altered zone due to friction during atmospheric flight, the Widmanstätten pattern being present centrally; a troilite nodule lies on the right.

(f) Supuhee H-group chondrite breccia of light-colored chondritic clasts set in a dark, fine-grained groundmass; PPL, $\times 46$.

(All specimens from the collection in the British Museum (Natural History).)

Iron meteorites (siderites)

The textures of iron meteorites are governed by phase relations in the Fe-Ni-P system (Romig and Goldstein, 1981), but this is no longer used as the basis for classification, which now depends on the abundances of Ni, Ga, Ge, and Ir (Dodd, 1981). The term *aerosiderite* is now little used.

Fe-Ni-P system. Below the melting point of Fe-Ni ($>1,450^\circ\text{C}$) the γ , face-centered cubic form, taenite, is stable. On slow cooling through 900°C pure Fe transforms directly to the α , body-centered cubic form, kamacite. Ni-bearing alloys enter the $\gamma + \alpha$ field at lower temperatures (Fig. 3). Here, with some undercooling, the low-Ni form nucleates within the high-Ni γ phase, equilibrium being maintained by lattice diffusion of Fe and Ni as temperature falls. The partitioning of Fe and Ni depends on both temperature and P content with small amounts of P stabilizing α and γ phases with higher and lower Ni, respectively, than in the P-free system. As temperature falls, lattice diffusion becomes increasingly sluggish and equilibrium can only be maintained in narrowing zones on

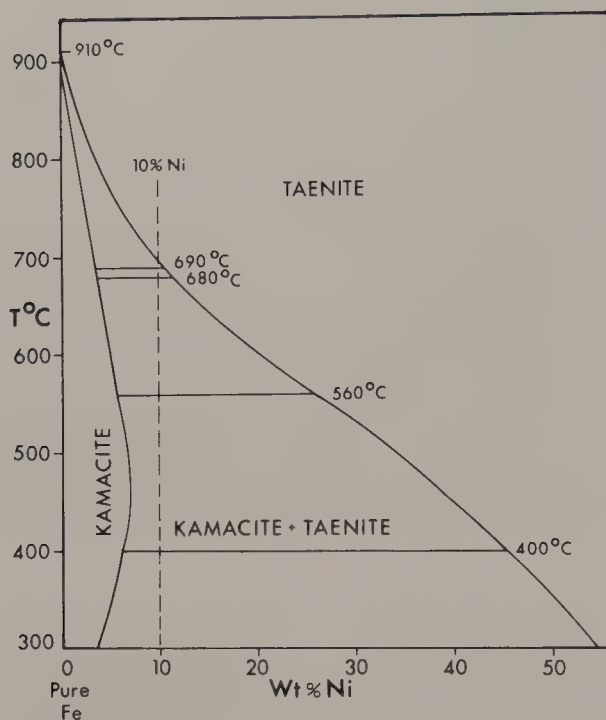


FIGURE 3. Fe-Ni system illustrating the Ni content of kamacite and taenite at different temperatures, metal with 10 wt% Ni, typical of octahedrites, occurs as γ Fe-Ni above 690°C; cooling below this temperature causes α kamacite to nucleate and grow as plates parallel to the {111} planes of the γ phase. At 680°C the kamacite begins to grow at the expense of γ -taenite, and at 560°C, the kamacite has grown to about three times the volume of the taenite, which has 25% Ni. Between 500 and 400°C the solubility of Ni in kamacite goes through a maximum, and at the latter temperature the kamacite sheds Ni as it continues to grow. By this stage at normal cooling rates the Ni builds up along the margins of residual taenite, the centers of which have not maintained equilibrium and have lower Ni contents than the margins. (After Hutchison, 1983).

either side of taenite-kamacite interfaces. Microprobe analyses of Ni content across such interfaces have been used to derive cooling rates in iron, chondritic, and stony-iron meteorites. However, the presence of P causes an increase in the diffusivity of Ni, so P content must be taken into account (Narayan and Goldstein, 1985).

Structures in iron meteorites. Structurally, iron meteorites may be subdivided into three types, although a number display anomalous structures resulting from unusual thermal histories (Buchwald, 1975).

1. *Hexahedrites* have low Ni contents (<6 wt.%) and consist of kamacite, schreibersite ((Fe, Ni)₃P) and a little troilite.

2. *Octahedrites* have intermediate Ni contents (6–ca. 15 wt.%). Cooling resulted in the nucleation and growth of plates of kamacite parallel to

the {111} “octahedral” planes of the parent taenite. On etching of a polished surface with acid the kamacite plates are revealed as up to four sets of parallel lamellae—the Widmanstätten pattern (Fig. 2e). Octahedrites are further subdivided into plessitic, finest, fine, medium, coarse, and coarsest types based on increasing width of kamacite lamellae, which roughly corresponds to decreasing Ni content. In fact, coarsest octahedrites have Ni contents as low as in some hexahedrites, but the former are poorer in P.

3. *Ataxites* are Ni-rich and in them kamacite did not nucleate along specific planes but occurs only as narrow intergrowths with taenite, called plessite, in randomly oriented zones.

Chemical classification. Thirteen groups of iron meteorites are recognized (Dodd, 1981; Kracher et al., 1980), but about 15% of iron meteorites have chemical compositions outside the confines of any group and are known as “anomalous” or “ungrouped.” The bulk chemistry of each group was presumably established by cosmic processes, but variation within each group was probably planetary in origin, caused by melting followed by fractional crystallization, or partial melting and segregation of a liquid. The members of two groups (IAB and IIE) typically contain silicate inclusions, and two other groups may be genetically related to stony-irons.

Stony-iron meteorites (siderolites)

1. *Mesosiderites* are polymict breccias of silicate fragments from the eucrite, diogenite, howardite association that were violently mixed with metal and related to the medium octahedrites of group IIIAB (Prinz et al., 1980; Dodd, 1981).

2. *Pallasites* are mixtures of olivine and metal. They lack evidence of shock and appear to have formed in a tranquil manner. Most (“main group”) pallasites are mixtures of metal related to group IIIAB and olivine of about Fo₈₈, but three (“Eagle Station trio”) comprise metal similar to group IIF and olivine of about Fo₈₀ (Kracher et al., 1980, and references therein).

Brecciated meteorites

Brecciation is common among meteorites. In rare cases, different classes of meteoritic material are mixed, for example, chondrite in aubrite (Cumberland Falls), eucrite in diogenite (Aioun el Atrouss), and H-group chondrite in LL-group (St. Mesmin). More commonly, brecciation occurred within a fairly uniform material (many H-group chondrites, Fig. 2f). In addition, more violent impacts produced shock-veins in some of which occur the high-pressure

polymorphs of olivine and pyroxene, and of ringwoodite and majorite (Dodd, 1981).

Other terms

The following terms occur in literature but are now little used.

Chladnite—a group name for achondrites essentially composed of orthopyroxene; the term originally applied to achondrites essentially composed of enstatite.

Cryptosiderite—a stony meteorite poor in Ni-Fe.

Grahamite—a type of mesosiderite.

Holosiderite—a meteorite consisting of metallic Fe without stony matter.

Ngavite—a chondrite composed of bronzite and olivine in a friable, breccia-like mass of chondrules.

Oligosiderite—a meteorite with only a small proportion of metallic Fe.

Sporadosiderite—a stony meteorite throughout which metallic Fe is disseminated.

Tadgerite—a black, semiglassy chondrite composed of bronzite and olivine.

Whitleyite—an aubrite (achondrite) containing fragments of black chondrite.

ROBERT HUTCHISON

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Cross-references: *Core; Lunar petrology; Shock metamorphism. Textures of igneous rocks; Ultramafic rocks.*

MID-OCEAN RIDGE AND OCEAN-FLOOR PETROLOGY

Over 70% of the Earth's surface is underlain by ocean crust, and through the mechanism of sea-floor spreading this crust was generated at the spreading centers located in the axial zones of the mid-ocean ridge system. Likewise, ocean crust is consumed and its constituents are recycled in the subduction zones beneath island arcs and cordilleran volcanic arcs. The whole cycle from creation to consumption appears to

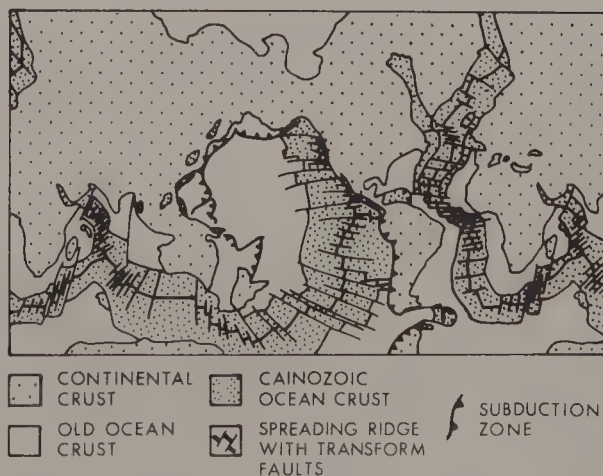


FIGURE 1. Distribution of ocean crust and principal active spreading ridges, with Cenozoic ocean crust (<65 Ma) shown.

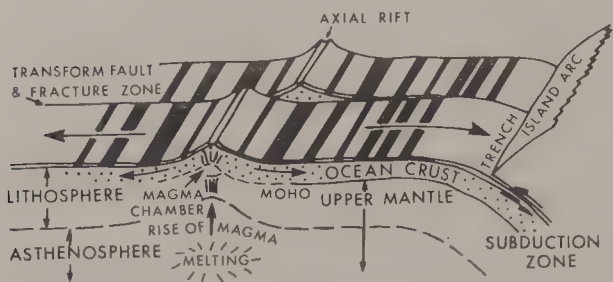


FIGURE 2. Schematic block diagram showing typical structure of ocean crust and the underlying upper mantle.

be completed in between 100 and 250 Ma; the oldest ocean crust identified in situ is only Jurassic in age, in contrast to the 3,800 Ma record preserved in the continental crust. The processes operative at mid-ocean ridges create ocean crust and impose its fundamental structure. As ocean crust moves away from these areas, it undergoes an aging process but retains the basic features produced by the genetic mechanism (Figs. 1,2).

Structure and petrology

Evidence for the nature of the ocean crust comes from four sources: (1) dredging, particularly on the mid-ocean ridges and the traversing valleys that follow transform faults; (2) the Deep Sea Drilling Project (DSDP); (3) geophysical surveys, particularly seismic and magnetic experiments; and (4) analogies with ophiolite complexes.

Mid-ocean ridges rise between 2,000 and 3,000 m above the abyssal plain and are usually crowned by a rift valley, or fault-bounded trough. The summit area is usually free of sediment, but a sediment drape increases in thickness away from the crest and may be over 500 m thick on the abyssal plain close to the continental slope.

The ridge is cut at right-angles by steep-sided valleys, whose depth may equal the abyssal plain to either side of the ridge. These valleys follow fundamental fracture zones, faults that displace the ridge crest.

Dredging of the deep ocean floor began in earnest with the HMS Challenger expedition in the 1870s, and basalt or spilite pillow lavas were recovered from the mid-ocean ridges at an early date. Sediment on the ridge flanks consists of calcareous oozes (pelagic chalk), but further away from the crest, in deeper water, this is replaced by radiolarian ooze (cherts) and terrigenous red lutites. The DSDP has demonstrated that this threefold division is not only spatial but vertical and secular on the ocean floor. Wherever the sediment cover has been penetrated, it is found to lie on basalt or spilite

pillow lavas, and the usual succession above begins with calcareous sediments followed by cherts and lutites.

The theory of sea-floor spreading was formulated to account for geophysical features, particularly the discovery of magnetic stripes of reversed polarity in the lava basement. These stripes run parallel to the ridge crest, and have a bilateral symmetrical arrangement about the axial zone. They are offset by the transform

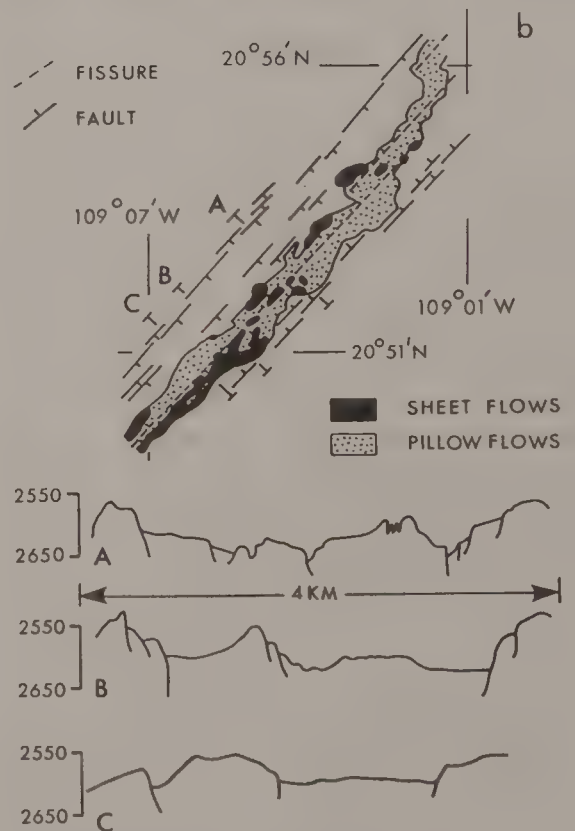
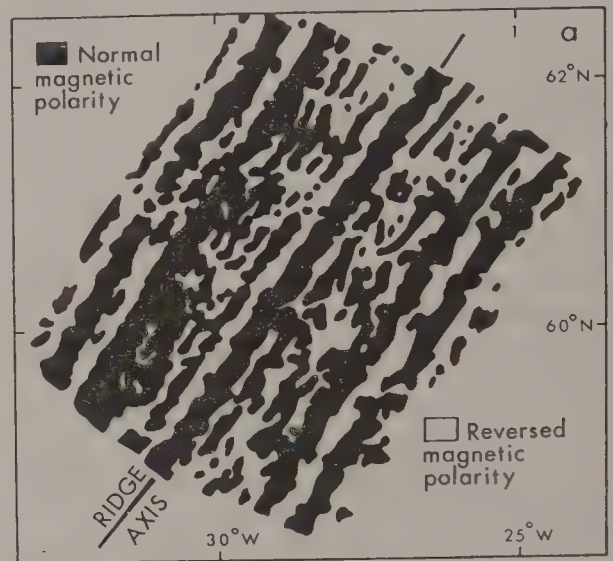


FIGURE 3. (a) Magnetic stripes on the ocean floor, Reykjanes Ridge, SW of Iceland. (b) Recent lava flows and topographic cross-section of the median valley of the E Pacific Rise. (After Ballard et al., 1981.)

fracture zones by the same amount as the axial zone of the ridge itself. The stripes themselves represent the orientation of the remnant induced magnetism of the lavas themselves, which records the polarity of the Earth's magnetic field at the time the lavas cooled through their Curie point (the temperature above which ferromagnetism is obliterated in a medium). As the Earth's magnetic field periodically reverses itself, the magnetic reversals on the sea floor represent a "magnetic tape" reflecting the eruption of lava at the ridge crest and the moving apart of the crust at the mid-ocean ridges. The youngest ocean crust is thus found at the ridge crest, and the oldest is

beneath those parts of the abyssal plain farthest from an active ridge (Figs. 3,4).

The presence of calcareous sediment on the ridge and siliceous sediments on the abyssal plain reflects the influence of water depth on the stability of calcium carbonate; active ridge crests may be shallow enough to protrude above the calcium carbonate compensation depth, so that calcareous ooze can accumulate. As the ocean crust moves away from the ridge, it sinks beneath this threshold so that calcium carbonate dissolves, or is precluded from accumulating, and its absence permits the slower accumulation of siliceous ooze.

Dredge hauls from the fracture zones,

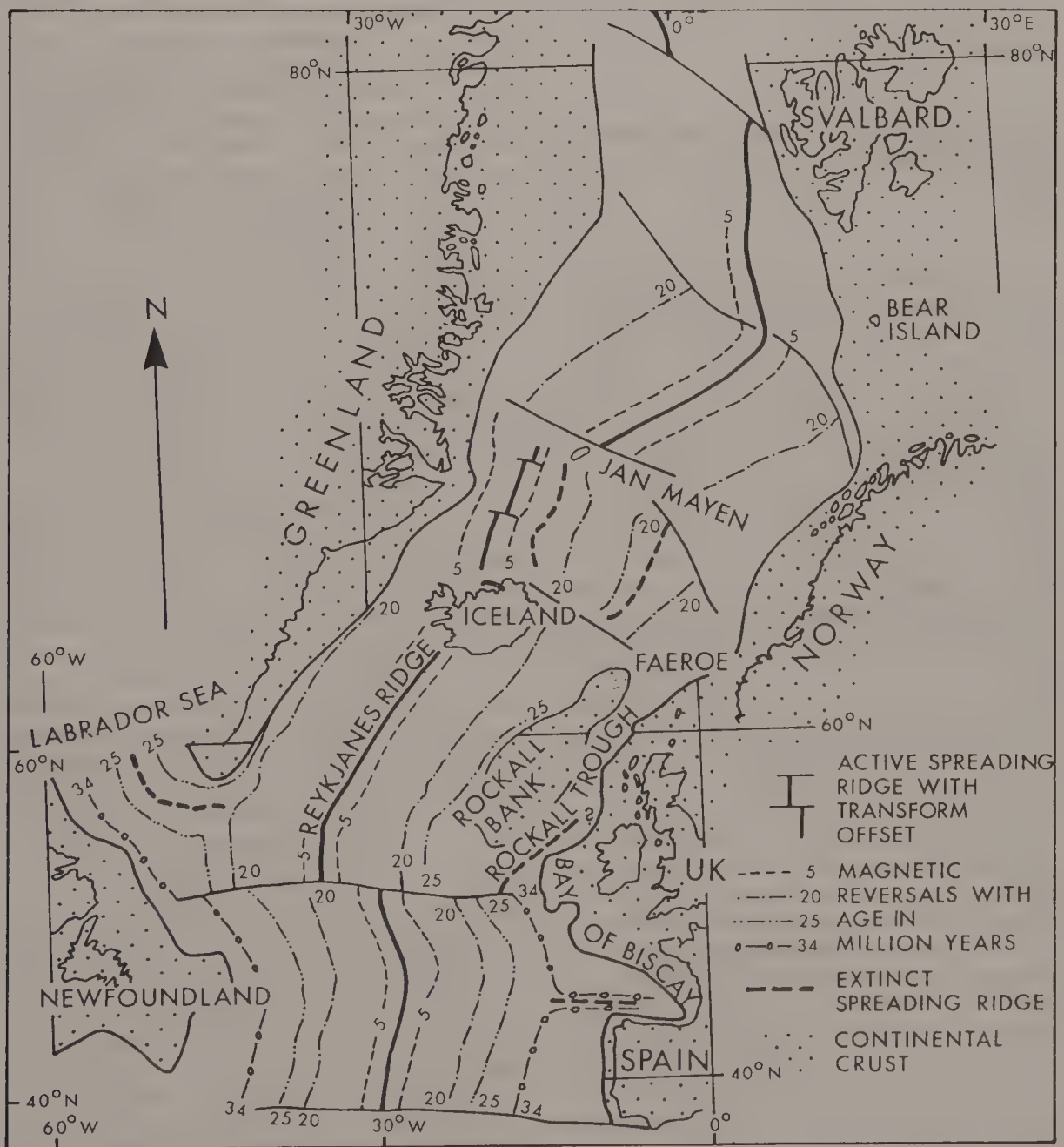


FIGURE 4. Ocean crust in the N Atlantic basin showing age of magnetic reversals and position of extinct spreading axes and transform faults. (After Talwani, 1978.)

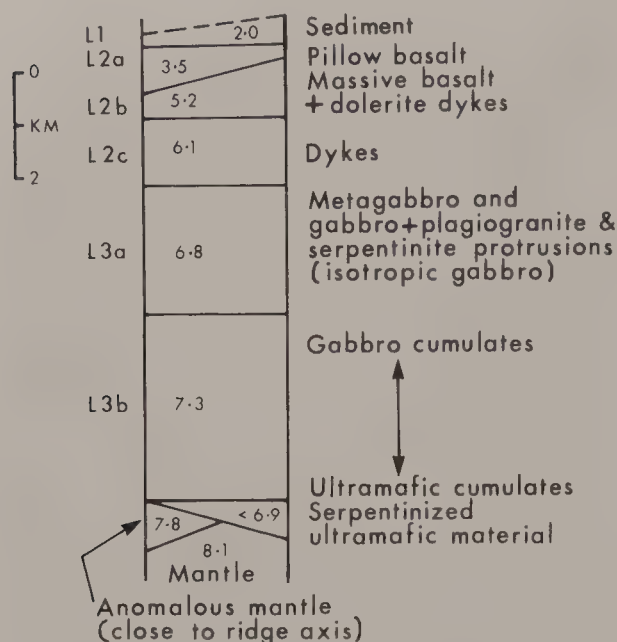


FIGURE 5. Schematic section through ocean crust showing the petrological layers (L) and geophysical signature (P-wave velocity, km/sec). (After Windley, 1984.)

particularly those that contain scarps, demonstrate the existence of other rock types in oceanic crust—dolerite, gabbro, diorite, trondhjemite, plagiogranite, anorthosite, various ultramafic rocks, serpentinite, foliated amphibolite, and flaser gabbro (gabbroic augen gneiss). At the same time, seismic profiles illustrate a layered structure with discrete P-wave velocities (Fig. 5). The deepest core samples recovered during DSDP have only sampled the upper part of layer 2, i.e., massive dolerite beneath pillow lavas and massive basaltic flows. However, the similarities between the seismic velocity layering, the structure of some ophiolites, and the similarities between rocks in ophiolite complexes and those dredged from fracture zones all indicate that some ophiolite complexes present representative samples of ocean crust. This permits modeling of the deeper ocean crustal layers.

On the basis of this approach, a model for the ocean crust suggests the following petrologic subdivision.

Layer 1. Sediments drape the older ocean crust. They consist of siliceous oozes with red terrigenous clays, overlaying a calcareous layer of variable thickness that may be absent altogether. Some lavas may be interlayered, and they are basalts or picrites, often spilitized and usually with pillow structures. They may be more Mg-rich and more mafic than typical ocean-floor basalts.

Layer 2. This layer consists of three subdivisions, all basaltic, that grade into each

other. Layer 2a consists entirely of tholeiitic pillow basalt and is often extensively spilitized. The interstitial cavities may be filled with calcareous pelagic chalk, palagonite, or chert, often hematite-rich or jasperitic. This grades down into layer 2b where massive basalt flows predominate and quartz dolerite (or diabase) dykes appear, some of which fed the basalt flows of this and layer 2a. In layer 2c quartz dolerite dykes eventually comprise the complete assemblage. Lava screens remain at the top, but at depth dyke is chilled against dyke. The lowest levels of layer 2c may be intruded by gabbro dykes.

Layer 3. This layer consists of gabbro and various gabbroic and ultramafic differentiates and cumulates. Layer 3a consists of quartz-bearing and quartz-normative gabbro or troctolite. The rock is massive with no layering or foliation (isotropic gabbro). There may be a petrologic gradation reflecting increasing SiO_2 content and falling $\text{CaO}:\text{Na}_2\text{O}$ ratio from troctolite, through quartz gabbro to quartz diorite and trondhjemite. These last two rocks form patches and layers in the isotropic gabbro. Dolerite dykes, autometamorphic amphibolite dykes and beerbachite dykes may be present. There is a gradation downward into layer 3b, which consists of gabbroic cumulates overlying ultramafic cumulates. The disappearance of feldspar marks the petrologically poorly defined "geophysical Moho." The base of layer 3b marks the base of the magmatic rocks in the ocean crust ("petrological Moho").

Layer 4. This so-called layer, which is the top of the upper mantle, consists entirely of ultramafic rocks; at the top these are usually harzburgite with pockets of dunite and wehrlite. This uppermost zone may also carry a foliation or mylonitic fabric. In ophiolites it is referred to as the ultramafic tectonite, although whether such a zone forms a continuous layer beneath layer 3 in the oceans is debatable. Samples of ultramafic mylonite have only been recovered by dredge haul from fracture zone scarps. The deeper levels of "layer 4" are essentially pristine mantle, consisting primarily of lherzolite.

Petrogenesis

The various layers of the ocean crust are considered to represent the generation, intrusion, differentiation, and eruption of a single magma type at the active spreading ridge. Essentially, melting in the upper mantle, which is depleted in large-ion lithophile (LIL) elements and the rare-earth elements (REE) when compared to the primordial mantle, generates a basaltic liquid that is SiO_2 -saturated or quartz-normative. This represents a 20–30%

partial melt of the mantle lherzolite. The melt migrates upward to accumulate in a magma chamber 3–5 km beneath the ridge axis. The residue from this melting, a mantle restite, is more Mg-rich than lherzolite, and is represented by the harzburgite-dominated zone, a residual layer. Processes within the magma chamber produce the cumulate zone, while the upper layer freezes against the roof as an isotropic mass (isotropic gabbro). Differentiation may produce sodic and siliceous by-products like quartz diorite and trondhjemite, but a large fraction is intruded into the rift zone to crystallize as quartz dolerite or erupt as tholeiitic basalt.

The magma chamber undergoes continuous replenishment and depletion, such that its overall bulk composition is maintained. This is reflected in the homogeneity of composition of all the components of layers 2 and 3. The thickness of the various sublayers in layers 2 and 3 appears to be related to the spreading rate and the rate at which magma is passed through from generation to eruption (Fig. 6a,b).

Transform faults and fracture zones

From place to place, spreading ridge axes are offset by transform faults (Fig. 7). As the active magma chamber is beneath the ridge axis, possibly at a depth of only 2 km, and the ridge is the site of lava eruption, some of the lava may spill into the adjacent fracture, which may be 1–2,000 m deeper. The presence of such relief provides talus into the fracture zone (Fig. 7a), but opposite the active ridge this will interdigitate with lavas—pillows or massive flows (Fig. 7b). At deep levels the intrusive activity has two effects: (1) dolerite and gabbro dykes may be injected into the fracture zone, or even the opposing crustal slab, and (2) a high-temperature aureole will develop in the relatively cool rocks of the fracture zone and the opposing wall. The wall rock and the rock-fill of the fracture zone undergo first a dynamic metamorphism, usually cataclasis, but this may be annealed within the superimposed aureole before undergoing further cool-working. A variety of banded amphibolites and gneissose rocks may be produced, intruded by dyke rocks, thermally metamorphosed and subsequently reworked. Such mylonites, banded amphibolites, gneissose, and flaser gabbros have all been dredged from oceanic fracture zones. These zones may also be the site of serpentinite (hydrated mantle material) diapirism. How high such diapirs rise is not certain, but dip-slip motion may assist in bringing them to the ocean floor, where they also contribute to the talus deposits. The nature of any pelagic sediment in such zones will depend on their proximity to the continental slope and the depth of the calcium carbonate compensation threshold (Fig. 7c,d).

Hydrothermal alteration and ocean-floor metamorphism

At the ridge crest, freshly erupted lava directly interacts with seawater, producing a variety of alteration products. Glass alters to palagonite very readily, and quench fractures permit contact between quite large volumes of water and rock even in massive flows. The presence of an active magma chamber as little as 2 km below the rift zone, and the deep faults bounding the rift, provide the heat engine and conduits for vigorous hydrothermal circulation. This has two consequences: (1) the igneous rocks of layer 2 are invariably hydrated and altered to varying degrees, usually in a well-defined series of zones or mineral facies, and (2) hydrothermal mineralization is ubiquitous. Where hydrothermal fluids debouch onto the ocean floor, often in a superheated state, massive precipitation of silica, sulfide and sulfate minerals results. Active hydrothermal

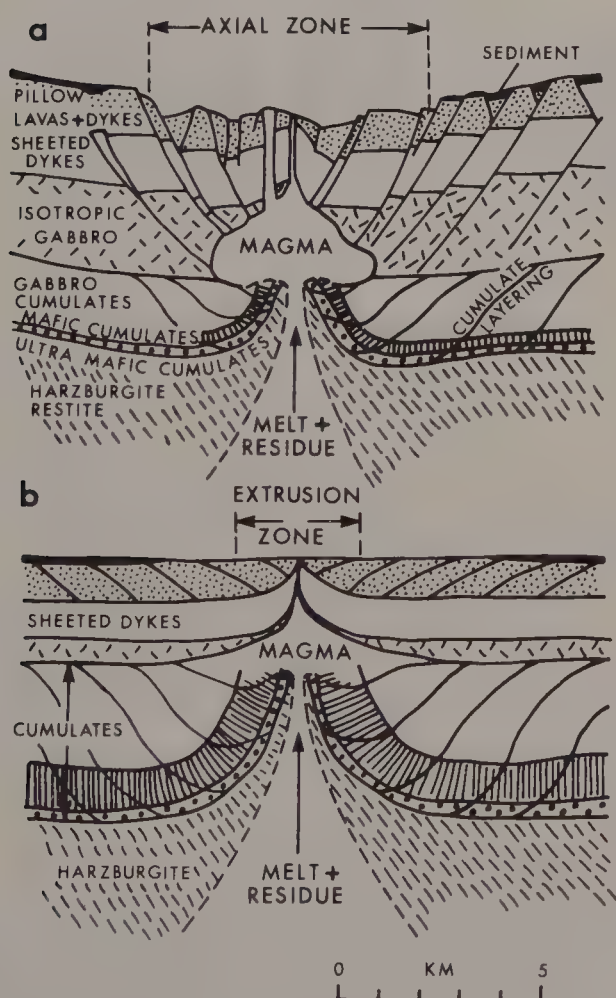


FIGURE 6. Structure beneath (a) a slow-spreading ridge axis, e.g., N Atlantic Ridge, (b) a fast-spreading ridge axis, e.g., E Pacific Rise. (After Windley, 1984.)

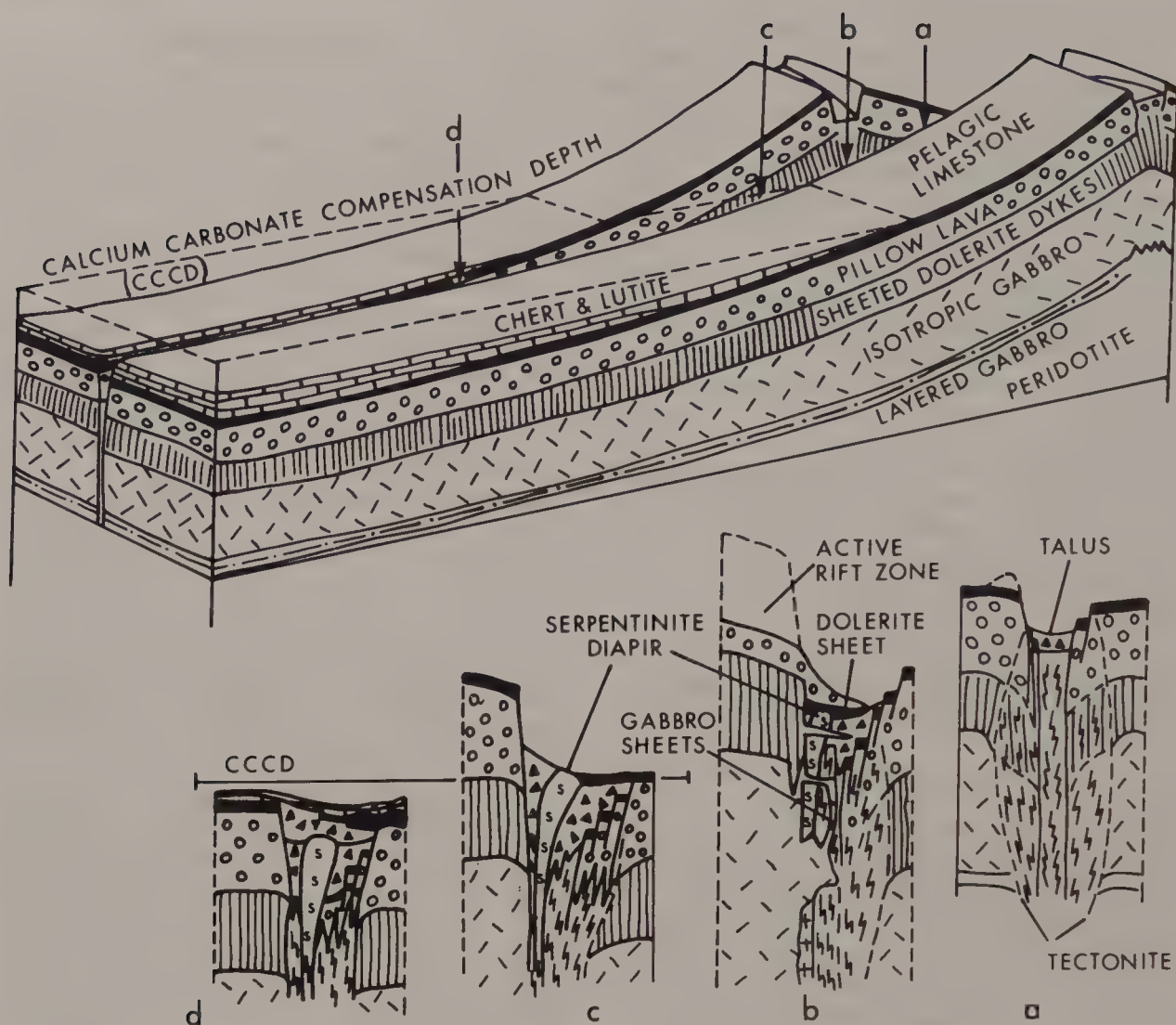


FIGURE 7. Schematic block diagram and representative sections through a spreading axis, mid-ocean rise, and offsetting transform fault-fracture zone. (After Karson and Dewey, 1978.)

mineralizing springs have been identified on the E Pacific Rise, in the Gulf of California, and on the Juan Da Fuca Ridge (off northwestern U.S.A.). Depending on the rate of terrigenous sedimentation this sea-floor mineralization gives rise to a spectrum of Cyprus- and Besshi-type sulfide deposits.

Beneath the sea floor, within layer 2, mineralization is generally disseminated and of the sulfide-stockwork type. This is accompanied by wall-rock alteration involving the replacement of mafic minerals by chlorite and the sericitization of feldspars. Zeolite growth, gypsum and carbonate growth are widespread, and much pillow basalt is altered to spilite.

ADRIAN F. PARK

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Cross-references: *Basalt*; *Hydrothermal processes*; *Igneous rocks—tectonic setting*; *Magmatic crystallization in open systems*; *Ocean-floor metamorphism*; *Ophiolite*; *Serpentinite*; *Spillite*; *Ultramafic rocks*.

MIGMATITE

Over wide areas of cratons and high-grade mobile belts, highly metamorphosed rocks are intimately mixed with granitic rocks. These composite rocks Sederholm (1907, p. 110, see Sederholm, 1967) called *migmatites* (from the Greek *migma*, mixture). They were originally defined as follows: "For the gneisses here in question, characteristic of which are two elements of different genetic value, one, a schistose sediment or foliated eruptive, the other, either formed by the resolution of material like the first or by an injection from without, the author proposes the name migmatities." Thus, migmatites consists of two component parts, a nongranitic host (usually schistose metasedimentary or metavolcanic rock, or gneiss) and a younger component that Sederholm's text implies has a granitic or granitoid composition. The two components are now commonly referred to as palaeosome and neosome, respectively, the latter also commonly referred to as a leucosome because most neosomes have a low color index.

Migmatites were divided by Sederholm (1913, 1926, see Sederholm, 1967) into two main groups, viz., *agmatites* or *intrusive breccias* (Fig. 1) and *veined gneisses* (*venites*) (Fig. 2) although there are transitional varieties. The intrusive origin of the veins was stressed and the resultant injection gneisses were called *arterites* (injection *venites*—Sederholm, 1926). In agmatite, the host takes the form of angular blocks disrupted by one or more sets of granitic veins. In the veined gneisses, granitic veins are commonly parallel or subparallel to the



FIGURE 1. Agmatite consisting of gray gneissose granite and amphibolite cut by veins of aplite; Spikarna, Tvärminne, Finland; scale bar = 50 cm.

preexisting planes of schistosity or foliation (cf., lit-par-lit injection of Michel-Levy, 1893) resulting in a banded rock (*stromatite* or *stromatolith*) with stromatolithic structure.

The veins themselves have a variety of compositions and textures, including the full range of granitoid mineralogy and textural types. More than one of agmatites and/or veined gneisses may be developed in any particular migmatite terrane. The relationship between vein and host may vary considerably from distinct, highly discordant "dykes" to less-discrete "blebs" and blind veins.

The nature and genesis of migmatites played a key role in the controversy surrounding the origin of granite, reviewed by Read (1957) and Mehnert (1968).

The study of migmatites has spawned a vast terminology, with terms used in relation to genesis, morphology, and process. In much literature, which aspect was implied has not been clearly stated and this has resulted in considerable confusion. In addition, most work on migmatites has concentrated on petrographic and geochemical aspects and has not dealt with structural control of much of the mixed aspect of migmatites (cf. Fig. 3g; Hopgood, 1984).

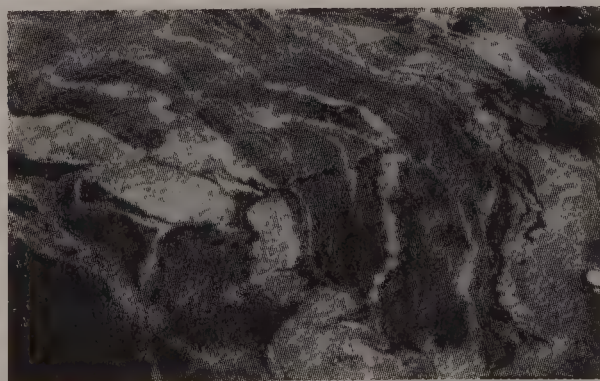


FIGURE 2. Veined gneiss (migmatite) consisting of hornblende schist and younger granite; Spikarna, Tvärminne, Finland; scale bar = 10 cm.



FIGURE 3. Types of migmatite: (a) agmatite, (b) dictyonite (diktyonite), (c) schollen-migmatite, (d) phlebite, (e) stromatolite, (f) surreitic migmatite, (g) migmatite with folds, (h) ptigmatic veins, (i) ophthalmic or augen structure, (j) stictolitic (fleck) structure, (k) schlieren structure, (l) nebulite. (After Mehnert, 1968.)

Descriptive terms

Leucosome—pale rock: the light-colored material that usually, though not always, is the neosome.

Melanosome—dark rock: the dark-colored material, often a biotite-rich selvage along the neosome–palaeosome boundary.

Mesosome—medium-colored material that often, though not exclusively, is the palaeosome; cf., Mehnert (1968) with Johannes and Gupta (1982).

Neosome—young rock, or vein material.

Palaeosome—early (old) rock, or host material to veins.

Morphological terms

Agmatite—originally referring to an intrusive breccia, but now used for a migmatite in which angular blocks of palaeosome are separated by either a meshwork of diverse, discordant neosome veins, or are surrounded by a net-veined complex of one neosome (Figs. 1, 3a).

Augen—see ophthalmic.

Chorismite—a megascopically composite rock; a little-used synonym of migmatite.

Diadysite—a network of granitic veins in a metamorphic palaeosome; generally similar to agmatite.

Dictyonite (*diktyonite*)—net-like and narrow, finely-spaced veins of neosome interlaced through the palaeosome (Fig. 3b); many veins follow fold hinge zones.

Dilational—see surreitic.

Fleck—see stictolithic.

Homophanous—a homogenous mass in which neosome and palaeosome are no longer distinguishable. Homophanous migmatite is strictly a contradiction in terms, but refers to Sederholm's conception of a progressive engulfing of the host by "granitic ichors" of which this type is the end result. Such a rock has no oriented fabric, and heterophanous migmatite could be used for all other migmatites. These terms should be used only when the transitional series proposed by Sederholm can be demonstrated.

Miarolithite—a chorismite having miarolitic cavities.

Metatect—the fluid or more mobile part of a migmatite formed as the result of anatexis.

Nebulite—a "cloud" migmatite in which the distinct boundaries between the different components are largely lost and often diffuse "ghosts" remain (Fig. 3l).

Ophthalmic or *augen* structure—contains either eye-shaped porphyroblasts of feldspar or aggregations of neosome with the same lenticular shape scattered through the palaeosome (ophthalmite—Fig. 3i).

Parautochthonous—refers to the mobile part of a migmatite that has moved, but not completely, out of its environment of formation.

Phlebite—a veined gneiss or migmatite in which irregular, but subparallel veins display folds and contortions (Fig. 3d).

Ptygmatic veins—neosome veins displaying fine-scale intricate and intense folding (Fig. 3h).

Schlieren structure—streaks of palaeosome or melanosome enclosing neosome; the streaks, or "schlieren" are often parallel to or define a foliation (Fig. 3k).

Schollen-migmatite—raft-structured migmatite resembling agmatite, but containing far more neosome in which palaeosome blocks are disrupted like mixed xenoliths (Fig. 3c); cf., *venite* and injection *venite* (*arterite*).

Stictolithic or *fleck* structure—structure consisting of aggregations of mafic material scattered through the palaeosome surrounded by pale haloes; poor in mafic minerals (Fig. 3j).

Stromatite (*stromatolith*)—layered veined migmatite or gneiss in which parallel veins or

neosome separate layered palaeosome, often parallel to the principal foliation or schistosity (Fig. 3e); cf., *epibolite* of Roques (1961).

Surreitic or *dilational* structures—structures in which neosome is located along lithological boundaries, in the cores of folds, or the necks of boudins (Fig. 3f).

Genetic terms

Anatexis—melting of pre-existing rock resulting in the development of in situ anatectic veins: an *anatextite* (*anatectite*) may have schlieren or nebulitic structure; an *embrechite* shows a regular parallel planar fabric.

Diatexis—high-grade anatexis, where melting has been complete or nearly so; a *diatextite* contains no continuous banding, i.e., it is homophanous, or nearly so.

Ectexis—in situ formation of a mobile part; the resultant rock is an *ectextite*.

Entexis—introduction from without of the more mobile part; the resultant rock is an *entextite*.

Metatexis—segregation of the mafic and felsic components of a rock, either by partial melting or by metamorphic segregation, into leucosome and melanosome.

Syntexis—melting of two or more lithologies to produce a hybrid melt, or the assimilation of palaeosome into the molten neosome.

Genesis

Sederholm (1907, 1913, 1926, see Sederholm, 1967) considered that the genesis of migmatites was intimately related to the activity of granitic "ichors," which were fluids gradational between aqueous solutions and a very diluted magma. These "ichors" (emanations), interpreted as emanating from granitic magma, were considered to have acted as solvents. Mobilization resulted from temperature increase associated with the rising magma. This new mobility imparted to rocks, resulting from anatexis, was referred to as *palingenesis* (Greek, "rebirth"); the product is referred to as a *mobilizate*. A different emphasis was given by Holmqvist (1907, 1921, see Read, 1957), who stressed the role of lateral secretion during ultrametamorphism, with the pegmatitic and granitic veins in migmatites being formed from bands in the schistose and gneisses rocks. A further matter of debate was the role of K-metasomatism in the migmatization process, with Härme (1965 and references therein) pointing out the role of structural control on the migration of potassium.

The continuing debate and the introduction of such concepts as a migmatite front, a basic front and a basic behind, and the significance of

migmatization (including *granitization* and *feldspathization*) in the genesis of granite are set out by Read (1957). While the genesis of the neosomes was predominant (see **Migmatite—origin of neosome**), the subsequent elucidation of the relationships of successively formed neosomes to successively formed structures has permitted not only further understanding of the development of the extremely wide range of migmatite “types,” but also (1) the use of neosomes as time-markers in isotopic studies (e.g., Hopgood et al., 1983) and (2) assessment of the evolution of migmatite complexes in the construction of plate tectonic models (e.g., Hopgood, 1984; see **Migmatite—structural relationships**).

D. R. BOWES

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MIGMATITE—ORIGIN OF NEOSOME

The “mixed” nature of migmatites inevitably means that at least two rock types must be considered in the reconstruction of the metamorphic history, or an appraisal of the orogenic significance of a migmatite. Cross-cutting field relationships between neosome and palaeosome (Fig. 1) may provide the initial evidence that the rock is a migmatite but alone does not indicate the origin of the neosome. In determining the origin of the neosome component, three factors must be assessed: (1) provenance (source rock), (2) process(es) operative, and (3) extent of movement. Independent assessment of these factors is not practical, as the composition of a source rock will have a bearing on the process(es) that may have been operative. The geothermal history of an area must also be examined in order to determine which process(es) had the greatest probability of producing neosome. It is only when the first two factors have been taken into consideration that the third can be addressed in the context of an overall geological history for an area where migmatite constitutes an important lithology.

The enigma of migmatite development in relation to granite genesis has been recognized since the introduction of the term migmatite (Sederholm, 1907, see Sederholm, 1967). Read's (1957) treatise *The Granite Controversy* focused on a historical consideration of the

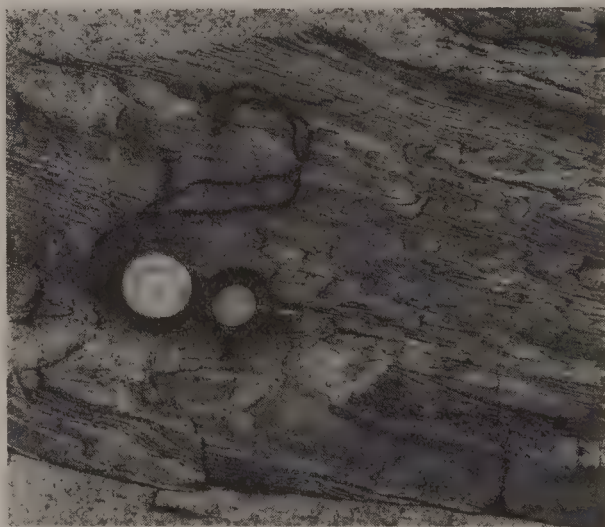


FIGURE 1. Migmatite composed of a mica schist palaeosome and two neosomes; fine quartzofeldspathic microlithons, probably formed by metamorphic differentiation and defining a gneissic fabric, constitute the first neosome; the second, granitic in composition and making up the thick light bands to the right and left of the photograph, cross-cuts the first; Savonranta, Finland.

problem of the origin of granitic fluids and their migration through rocks producing a progressive, and bulk, transformation of one rock into another, specifically with reference to the production of granites. White (1966) listed four processes that were considered to give rise to migmatites of granitic character: (1) bodily injection of granitic magma from an adjacent igneous intrusion; (2) production of the granitic neosome portion by metasomatism; (3) partial melting and subsequent segregation of the lowest melting point constituents during high-grade regional metamorphism; (4) segregation without melting—metamorphic differentiation. However, no hard and fast boundaries can be drawn between the processes simply on the basis of the macroscopic aspect of migmatites (limiting the effectiveness of field work in discerning the origin of neosome), especially where two or more processes could have contributed to neosome production. No rock is ever a closed system during metamorphism relative to pressure, temperature, and chemistry, so the four processes must be viewed as end points in a continuous spectrum; strictly speaking, this precludes isochemical metamorphism as a process that could give rise to neosome.

Mehnert (1968) classified migmatites on the basis of the gross relationships between neosome and palaeosome and also suggested that cross-cutting, sharply-defined neosome contacts were indicative of the neosome being allochthonous (Figs. 1, 2). In the same way, diffuse contacts could be indicative of metasomatism, metamorphic differentiation, or an origin by *in situ* partial melting. Excluding metasomatism, which definitely implies the introduction of an exotic component(s), the other two processes could give rise to autochthonous neosome given suitable conditions of pressure and temperature.

The ambiguity inherent in the subjective interpretation of neosome contact relationships requires that a more quantitative approach be used to elucidate neosome development, the most effective tools being petrographic and geochemical analysis. The petrography of a rock reflects its bulk chemical composition and so may act as a first-order discriminatory feature in deciding whether there is a genetic link between a neosome and palaeosome. For example, where granitic rocks intrude granulites or basic rocks, there is little chance of *in situ* partial melting having given rise to the neosome; the mineralogy and chemistry are sufficiently different (e.g., the lack of K-feldspar in the palaeosome) as to necessitate an allochthonous origin for the neosome. In a relationship like that described, the temperature of the neosome



FIGURE 2. Migmatite composed of a granodioritic palaeosome and thick cross-cutting veins of foliated granitic neosome that also contains large potassium feldspar phenocrysts; emplacement into a coarse equigranular granodioritic rock, sharply defined contacts, and the clear mineralogical and textural differences are indicative of an allochthonous derivation for the neosome; Savonranta, Finland.

(probably ca. 700°C) will be too low to supply sufficient heat to melt phases in the palaeosome, whose melting temperature exceeds 900°C. This is a direct measure of the specific heat capacity of the neosome compared with the latent heat of fusion of the palaeosome. A more typical relationship in migmatitic terrane is that of a palaeosome of pelitic to psammitic schists and a quartzofeldspathic neosome in the form of veins (Fig. 1). The major components of both the neosome and palaeosome in these cases are SiO_2 , Al_2O_3 , K_2O , Na_2O , CaO , FeO , MgO , and the most significant phases in both neosome and palaeosome are quartz, K-feldspar, plagioclase, and biotite. Clearly, in this case an *in situ* derivation is possible on the basis of mineralogy, but may be accompanied by an unknown amount of neosome redistribution; in Precambrian migmatitic terranes such as those of Greenland and the Baltic Shield, the latter factor could be of considerable importance.

Analysis of the cryptic variations or trends in the geochemistry of the neosome and palaeosome (as well as the use of absolute geochemical constraints, e.g., Sr and Rb isotopes) in cases where no mineralogical distinction can be drawn, becomes the most powerful tool in determining the origin of a neosome and in assessing how far it is likely to have moved.

Assessment of the four processes proposed by White (1966) can be made in relation to those geochemical and/or physical features that are compatible or incompatible with that mode of neosome origin, assuming a generalized migmatite composed of a semipelitic schistose palaeosome and quartzofeldspathic vein neosome.

In the case of partial melting, consideration of the normative proportions of Ab, An, Or, and Q in the neosome in relation to the granite system tetrahedron (Fig. 3) will indicate whether a granitic neosome represents a minimum melt composition (cf. Winkler et al., 1975). To derive a melt of this composition, the palaeosome schist would have to be heated to ca. 640°C under conditions of moderately high P_{H_2O} , at ca. 5 kb total pressure. Problems exist in incorporating the 7–14% water (by weight) in the granitic melt that this water pressure would imply in a natural system. In a coexisting semipelitic palaeosome, the presence of phases like garnet and sillimanite would provide further corroborative evidence for a phase of ultrametamorphism concomitant with partial melting. Allied with observations in the field that the neosome is present as “blind” veins with diffuse

contacts (loosely referred to as resulting from “sweating out”) against a more mafic palaeosome (sometimes referred to as a “complementary mafic selvage”), an in situ derivation of the neosome would be a satisfactory conclusion if the neosome had a eutectic composition.

In a case where the neosome represented a minimum melt, the palaeosome contained no phases consistent with a high grade of metamorphism and there were sharp contacts, the implication would be that the neosome had been emplaced subsequent to derivation elsewhere. Isotopic constraints in the form of initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios will indicate, in the case of disparate neosome veins, whether they belong to the same isotopic system, i.e., crystallized from the same melt. The same reasoning may also indicate a genetic association with (or distinction from) any nearby granitic intrusion. Once the age of the rock is established, then the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the neosome will indicate, if it is high, that the neosome had been derived by partially melting a rock with a previous crustal history; if the initial ratio is low it might indicate derivation of the neosome by differentiation of a magma separated from the mantle, or by partially melting a source that had been a recent crustal addition. Reasoning of this type, integrated with Pb isotopic data has been used to elucidate the petrogenetic history of the Amîtsoq and Nûk gneisses and Qôrqt granite of W Greenland (Moorbath et al., 1981). For this approach to be successful, and to gain an insight as to whether the neosome had been emplaced from below or had been derived laterally (suggesting that partially melted rocks would be present at the same crustal level), it requires integration with comprehensive regional mapping and geochemical analyses.

If, for the granitic neosome being considered, partial melting is a tenable hypothesis, then there are three further lines of approach that may help constrain a possible source. (1) If the ratio of $\text{Al}/(\text{Na} + \text{K} + (\text{Ca}/2))$ is >1.1 , indicating a peraluminous association, then this would be compatible with derivation from a semipelitic to pelitic source (Shand, 1950). (2) It might be possible to refine a genetic association between a neosome and a palaeosome if they lie on the same “unmixing line” in a variation diagram (Fig. 4; White and Chappell, 1977); any parent melt derived, probably differentiated by crystal fractionation. (3) Tie lines between coexisting biotite and K-feldspar in the ternary system defined by Al_2O_3 , K_2O , and SiO_2 would, for juxtaposed neosome and palaeosome, be parallel if the coexisting phases had recrystallized under similar conditions, as might be expected during in situ anatexis. If the tie lines crossed, then it might be inferred that

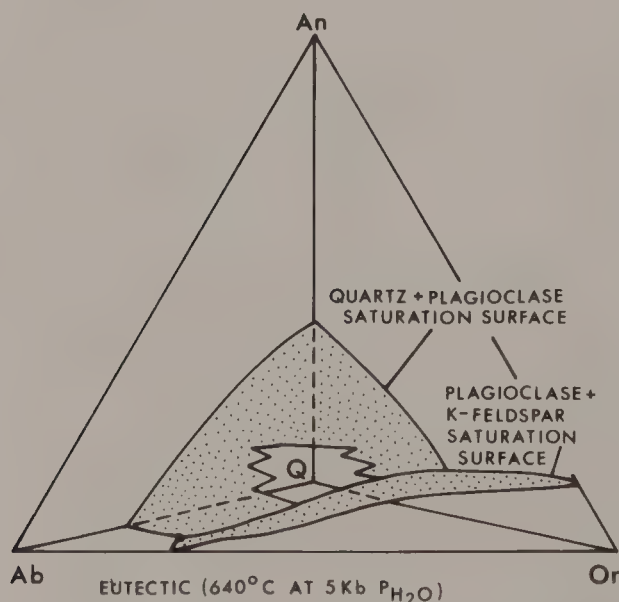


FIGURE 3. The granite tetrahedron (cf., Winkler et al., 1975). Q, quartz; Ab, albite; An, anorthite; Or, orthoclase.

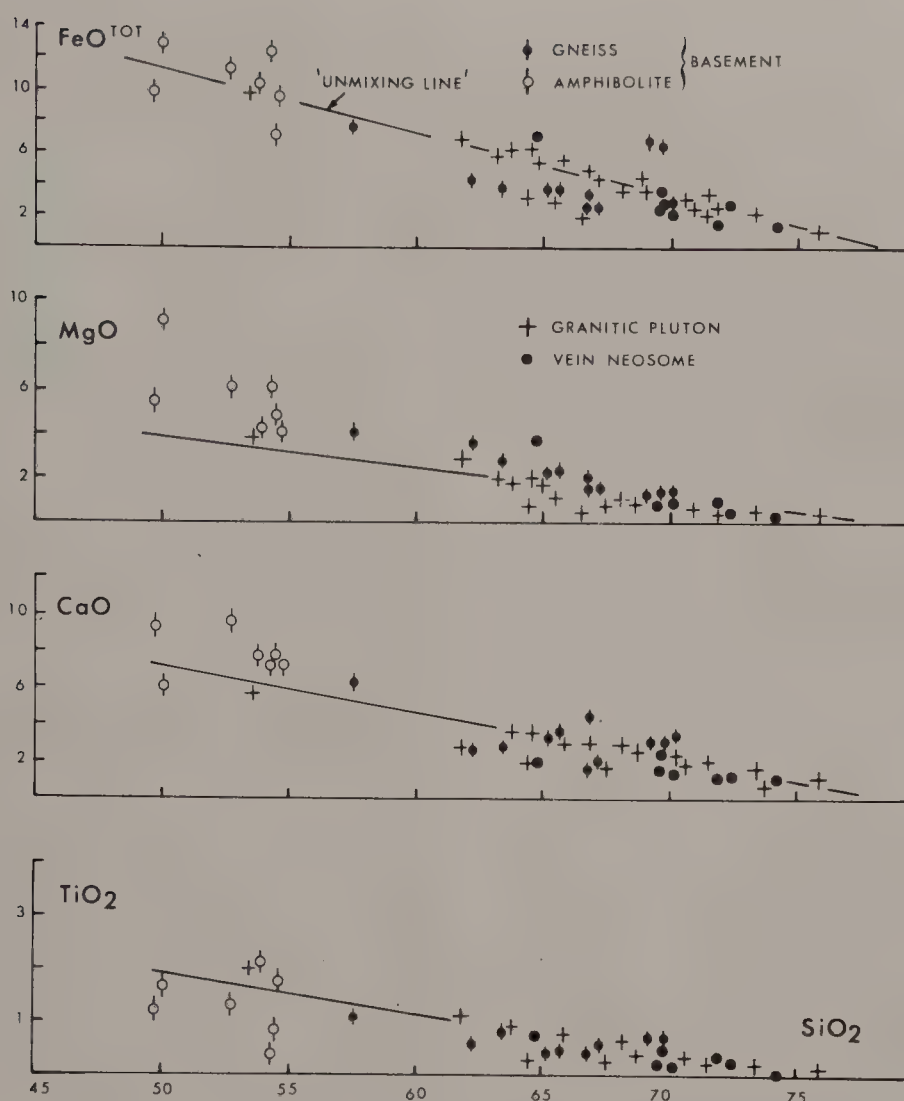


FIGURE 4. "Unmixing lines" demonstrating a possible genetic relationship between granitic plutonic rocks and associated vein neosome and pointing to the former being derived from the latter; the progressive change represented by the trend within the granitic rocks is probably a function of crystal fractionation; Svecokarellides of E Finland.

equilibrium conditions of crystallization were different, as would be expected if the neosome were allochthonous.

A number of considerations follow if the neosome does not represent a eutectic melt within the granite system: (1) the quartz-feldspar, and feldspar saturation surfaces defined by Winkler et al. (1975) at 5 kb $P_{\text{H}_2\text{O}}$ cannot be used to define neosome composition derived under different conditions of water pressure; (2) the neosome composition may be biased away from the eutectic composition because of depletion of one or more of the granite components in the source; (3) there may have been fractional crystallization of the neosome melt accompanied by preferential depletion or enrichment of a certain phase; (4) another process was responsible for the neosome production, e.g., metamorphic differentiation may be a more compatible mode

of origin for neosome veins consisting solely of quartz and plagioclase.

Many migmatite terranes are typified by schistose palaeosome interlayered with many parallel, leucocratic quartzofeldspathic neosome veins, trondhjemitic or plagiogranitic in character, that lack any direct igneous association that could be used to invoke a crystal fractionation origin. An alternative hypothesis for the production of this kind of neosome is that metamorphic differentiation was operative. This process, which more or less by consensus in the literature, results in the redistribution of quartzofeldspathic material during prograde metamorphism, probably starts with the creation of a spaced cleavage at lower greenschist facies and ends with partial melting becoming the dominant process at upper amphibolite facies. Rather than the breakdown of crystal structure typical of partial melting, metamor-

phic differentiation involves a "solid state" transference of ions (e.g., Na^+ , K^+ , Ca^{2+} , Al^{3+}), possibly in the form of an intergranular aqueous fluid migrating along grain boundaries. Excluding the introduction of material from an extraneous source, metamorphic differentiation might be taken as synonymous with isochemical metamorphism.

The presence of water in quartz (ca. 100 p.p.m. hydrogen) results in a hydrolytic weakening of the quartz structure, making it more susceptible to ductile deformation and redistribution in the solid state (cf., Glen,

1982). The concentration of this material in zones of reduced pressure or low strain contributes to the interlayered aspect of quartzofeldspathic and mafic layers typical of many migmatites. The concentration of water in particular horizons may act as a precursor or catalyst during this kind of redistribution. This situation is distinct from the dissolution of quartz in the form of an aqueous solution, where quartz would have a solubility of 1% at 550°C –4 kb in water (conditions consistent with mid-amphibolite facies), the solubility increasing in a more acidic solution. The behavior of plagioclase during this kind of metamorphism is likely to be similar to that of quartz; having a three-dimensional crystal structure, the incorporation of water will again result in hydrolytic weakening. In plagioclase, the replacement of Na^+ by H^+ ions (the latter being much smaller) will also weaken the plagioclase structure. This effect is likely to be directly proportional to the pH of any coexisting aqueous solution.

Grain boundaries are the most likely locus for the redistribution of this material, as flaws and fractures will be concentrated in these regions acting as a form of incipient permeability and porosity. Qualitative support for this is shown in Fig. 5, indicating that the grain boundaries between coexisting plagioclase crystals coincide with anomalous concentrations of certain elements of a scale significantly different from any zoning. In Fig. 5a there is a sudden drop in the CaO and Na_2O contents of the coexisting plagioclase crystals at their mutual boundary.

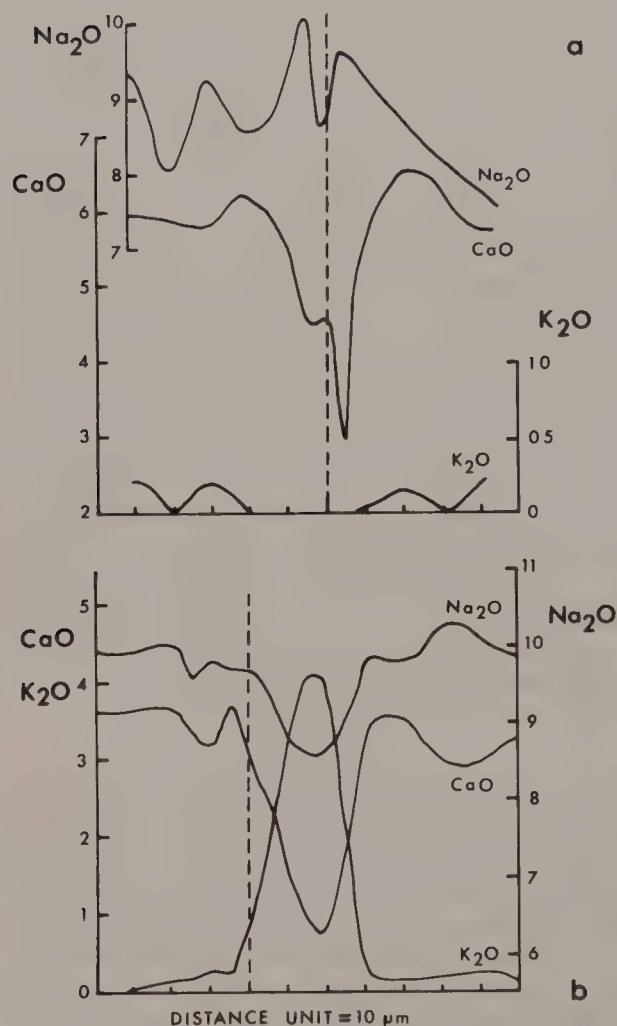


FIGURE 5. Element concentrations at grain boundaries. (a) Electron-microprobe traverse of a grain boundary (optical grain boundary observed at the dotted line) between two plagioclase crystals; the coincidence of the CaO and Na_2O narrow troughs approximating to the grain boundary suggests that both constituents have been removed together, with metamorphic differentiation the operative process. (b) Traverse of grain boundary between two plagioclase crystals, similar to (a); the coincidence of the CaO and Na_2O suggests their simultaneous removal, while the K_2O peak suggests the introduction of K, with metasomatism the operative process.



FIGURE 6. An enclave of mica schist surrounded by granite, close to the granite–country rock contact; the mica schist shows a biotite selvage, suggesting removal of quartz and feldspar and the development of K-feldspar porphyroblasts within the biotite–quartz–plagioclase matrix of the schist.

The scale of the decrease in terms of both lateral range and wt.% oxide, and the fact that the Na and Ca troughs coincide, indicate that this is not zoning in the accepted sense; it may be inferred from this plot that Na and Ca were removed from the crystals and presumably redistributed elsewhere. Although only three variables were plotted, there is no indication of an introduction of K, suggesting that, whatever process was removing the material, it at least approximated to isochemical metamorphism. A different, yet similar situation is indicated in Fig. 5b where, coinciding with the drop in CaO and Na₂O, there is a concomitant increase in K₂O, consistent with the removal of Ca and Na and the introduction of K; this may be representative of K-metasomatism.

The creation of mafic selvages during metamorphic differentiation is a feature that the

process has in common with partial melting (Fig. 6). A possible method of distinction between two neosomes would be the electron microprobe analysis of plagioclase crystals within the veins and the host palaeosome. Figure 7 shows the results of such a traverse from a vein with a clear selvage and another with a sharp contact. The former shows that even after recrystallization the composition of the plagioclase is consistent with that in the adjacent host mica schist. This suggests that even after a clear separation of the quartzofeldspathic portion there is little difference in the An content of the plagioclase, which is a situation compatible with metamorphic differentiation or isochemical metamorphism (Mehnert and Busch, 1982). Any difference in the An content of plagioclase is insignificant when compared with what would be expected

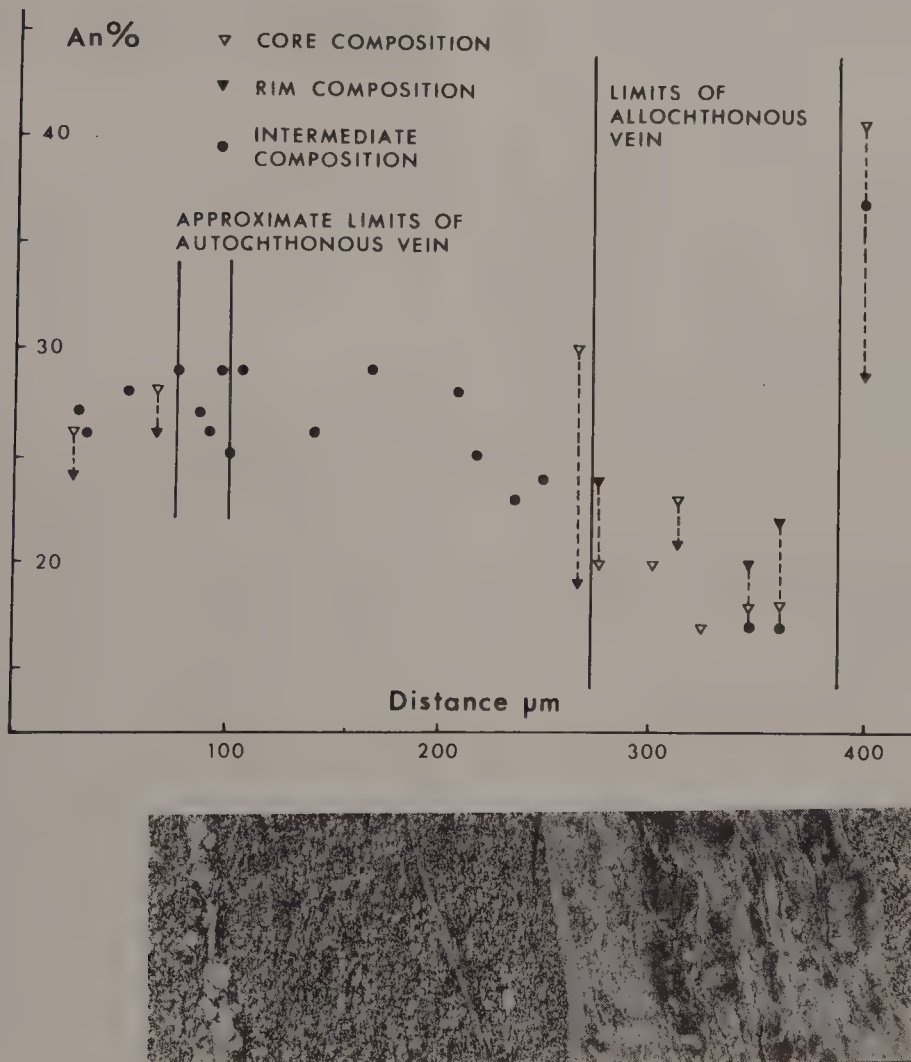


FIGURE 7. Plot (corresponding to photograph) of the An content of plagioclase crystals from coexisting, cross-cutting neosome (allochthonous vein), mica schist and a supposedly metamorphically differentiated neosome with a biotite selvage (autochthonous vein); An content of the plagioclase from the schist and the autochthonous vein coincides, whilst that recorded for the allochthonous vein is noticeably lower.

from the melting of plagioclase, considering the shape of the liquidus in the albite–anorthite system. The vein with the sharper contacts is clearly typified by plagioclase compositions that are consistently more albitic; this would be expected in a partial melt situation accompanied by the removal of the liquid phase. This presumes that the plagioclase in the source was at least more An-rich than that presently in the vein and probably more An-rich than that in the schist. More calcic rims to the plagioclase in the veins with the sharp contacts, allied with the Ca depletion in the plagioclase in the adjacent mica schist, suggests a later metamorphic reequilibration.

Metasomatism can be referred to as a metamorphic change induced in a rock by the introduction of material from an external source. The extraneous material is often considered to be in the form of an aqueous or volatile-rich solution ultimately responsible for the bulk transformation of a host rock. The process is commonly referred to as *feldspathization* or *granitization* in relation to migmatites (cf., Read, 1957) where the incoming fluid would be expected to be K-rich. The macroscopic aspect of metasomatism, in the case where the altering fluid is “granitic” in character, is often expressed as the development of feldspar phenocrysts within a matrix that has a progressively increasing proportion of quartzofeldspathic constituents (Figs. 6, 8). In most instances the *bulk* transformation of a rock precludes the creation of a macroscopically heterogeneous rock (i.e., a migmatite consisting of a neosome and palaeosome), unless some other agency is involved (e.g., directed pressure).

Fluids generally considered to be the altering agent in metasomatism are often highly differentiated liquids rich in volatile constituents, alkali earths, and other LIL elements. In an effort to elucidate the origins of a fluid of this nature, analysis of the $^{18}\text{O}/^{16}\text{O}$ ratio will indicate whether the aqueous component of the fluid was of magmatic, meteoric, or metamorphic origin, the latter probably produced by dehydration reactions at high grades of metamorphism (cf., Cox et al., 1979).

As a process giving rise to neosome introduction, igneous injection is normally associated with sharp contacts and a clearly definable mineralogical and/or textural difference between the neosome and palaeosome. Criteria by which to judge the origin of such a neosome would be the same as those applied to a partially melted or metamorphically differentiated neosome.

Recent work concerning the origin of neosomes has included detailed structural



FIGURE 8. A quartz diorite having undergone the gradual process of granitization; there is no clearly defined neosome and palaeosome, yet separate areas of different texture and mineral constituents with diffuse boundaries are indicative of a gradual transformation; Savonranta, Finland.

analysis that allows a time-relative neosome stratigraphy to be developed in migmatitic terranes (Hopgood and Bowes, 1978; Hopgood, 1984—see **Migmatite—structural relationships**). When integrated with detailed isotopic analysis, this not only allows the application of a radiometric framework (Hopgood et al., 1983) to a migmatite, but also may indicate changes in the source region for a sequence of neosomes, or allow progressive changes in neosome geochemistry to be assessed.

Analysis of the fluid inclusions in neosomes can provide evidence for the nature or the changing nature of the volatiles present during the crystallization of a neosome (Yardley, 1983) and may be used to determine the crustal level at which neosomes developed. One of the more difficult problems to be tackled is the determination of the absolute volume of neosome that has pervaded a rock (cf., Olsen, 1983). The introduction of neosome will destroy any porosity that a rock may have had, and analysis of the neosome may not include the most highly differentiated liquids, which may have been lost.

A form of migmatite rarely considered when addressing the problem of the origin of neosome is the migmatite created by the deformation of an interlayered lithological sequence within which there are distinct competence contrasts

resulting in varying responses to ductile deformation; in this situation there are no neosome and palaeosome *sensu stricto*. Fluid movement and emplacement may accompany such a process but need not be responsible for the development of the initial migmatitic aspect of the rock (Halden et al., 1982).

N. M. HALDEN

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Cross-references: *Anatexis*; *Eutectic melting*; *Granite*; *Metasomatism*; *Migmatite*; *Migmatite—structural relationships*.

MIGMATITE—STRUCTURAL RELATIONSHIPS

The complexity of migmatites is such that some workers have regarded their structural development in a different light from that of other tectonites. This has led to interpretations of response to stress that differ significantly from that for lower-grade metamorphic rocks, with structures of migmatites being considered to be irregular to a degree that is haphazard or “wild”—but see Berthelsen et al. (1962) on the so-called wild migmatites. Such an interpretation stems, at least in part, from evidence indicating the presence of melt at some stage(s) during migmatite development. Although partial melting forms an integral part of much migmatite formation, the presence of patches of fluid material, or of extremely ductile material, does not preclude the pervasive development of penetrative structures. Observation shows that structural sets form in migmatites just as they do in other, lower-grade metamorphic tectonites and that refolding relationships between sets can be determined in these rocks as in other tectonites, as can cross-cutting relationships of superimposed planar fabric elements. The explanation for the apparent paradox of pervasive fold structural sets in rocks containing melt material is a simple one. It is that the melt is developed irregularly, and often discontinuously, rather than as a continuum. Even if the pockets of melt are interconnected, the relationship between the melt and the host tectonite is one of fluid between septa rather than disconnected rafts in a fluid host, so that the stress can be transmitted throughout the rock mass. Only in the case where there are isolated rafts in a melt would the stress fail to be

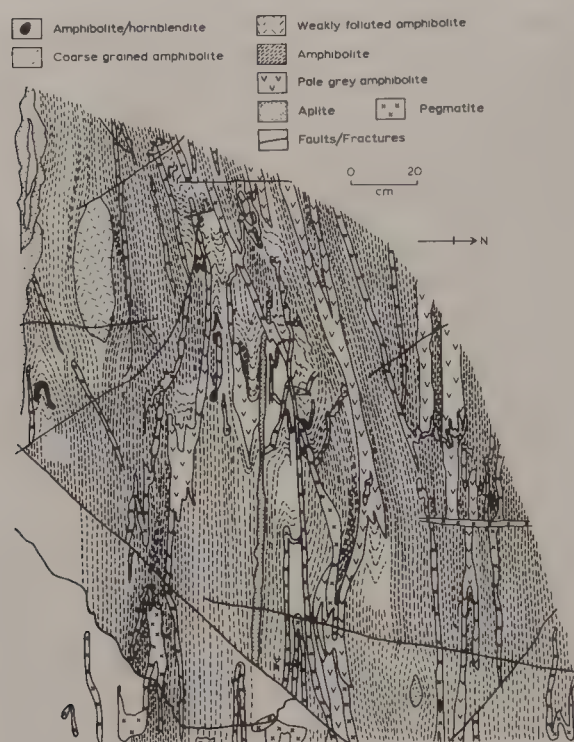


FIGURE 1. Relationships between veins, early folds, and fractures in migmatite showing the "mixed" (i.e., migmatitic) aspect of the rocks; Proterozoic Svecof Karelian migmatites, Skåldö, S Finland. (From Hopgood et al., 1976.)

transmitted throughout the rock mass to produce a pervasive structural pattern.

Some observers have found the apparent structural complexity in migmatites (Fig. 1) to be such that its resolution has not been possible. However, such apparently irresolvable "wildness" does not seem to be generally characteristic of migmatite complexes, as has been demonstrated in Finland (Hopgood, 1984), Scotland (Hopgood and Bowes, 1972), India (Halden et al., 1982), and elsewhere. Two basic principles underlie the elucidation of these structural relationships, viz.: (1) the relative age of structures can be determined on the basis that later structures deform those earlier in the sequence (Figs. 3, 5), and, similarly, discordant fabrics and veins are younger than the features they cut or affect (Hopgood, 1980); (2) structures can be correlated in much the same way as lithological units can in stratigraphy. If, for example, of units (i.e., structures) A, B, and C at some particular locality, B and C can be correlated with B and C of structures B, C, D, and E, say, at a second locality and C with C of structures C and F, say, at a third locality, then an overall correlation is made. The difference is that in structural correlation the parameters, such as lithology and palaeontology of stratigraphical correlation are replaced by "style." Style, a term whose meaning is commonly

misunderstood, embodies in its broadest sense *all* the features of a particular structure; these include features such as limb thickness, hinge shape, axial planar features, associated mineral growth (and thus *P-T* conditions of development) and orientation, and not solely the geometry. This means that the characteristic features of a fold of any particular set, or "phase," taken together, are distinctive and very often unique. The fold sets of different relative ages in any exposure together constitute the local partial succession and the partial successions from different exposures can be correlated and integrated to build up the full succession for the region. Furthermore, the style of these fold sets in their particular relative order comprise the overall character (total style) of this specific succession and thus enable it to be distinguished (in a manner analogous to the distinction of a particular stratigraphical succession) from any other, superficially similar, succession elsewhere, even though it has, or may have, the same number of fold sets.

Structural relationships in migmatites of southern Finland

Structural relationships in migmatites are probably nowhere better seen than in the magnificently exposed early Proterozoic Svecof Karelides in the archipelago of southern Finland, the subject of J. J. Sederholm's work (Sederholm, 1967) at the turn of the century. It was he who proposed the term "migmatite" to describe the mixed aspect of these rocks, which may be taken as typical of migmatite terranes in general both in terms of the structures they show and the rock types present. They display an intricate mixture of supracrustal rocks (micaceous, quartzofeldspathic, and hornblende gneisses, amphibolites, and marbles) together with a wide variety of plutonic and hypabyssal rocks (metagabbro, metadiorite, quartz diorite, granodiorite, granite, and acidic pegmatites). Ultrametamorphism and anatexis have resulted in the formation of leucocratic neosome, much of it of tonalitic-granodioritic composition. Throughout their history there has been repeated deformation and this is shown by the development of several sets of structures (Fig. 2). The emplacement of syntectonic veins adds to the lithological complexity with successively emplaced veins having cut across successively formed folds and having been deformed by subsequently developed folds. As deformation progressed, the early forms of folds produced by buckling, flexure or slip have been considerably modified by movement on widely separated discrete fractures parallel to the fold axial planes, and by flow as well as by dislocation

Fold set	Fold structure	Fold axial trend	Fold description	Neosome	Rock type	Dating method	Age (Ma)
					LAVAS: VOLCANOGENIC, DETRITAL AND OTHER SEDIMENTS	U-PB ZIRCON	1895
F ₁		E-W	INTRAFOLIAL ISOCLINAL FOLDS		MAFIC (GABBROIC) PEGMATOIDS QUARTZOFELDSPATHIC VEINS AMPHIBOLITE FACIES PARA-GNEISSES	U-PB ZIRCON U-PB MONAZITE	1885 1875
F ₂		E-W	TIGHT FOLDS		PERVASIVE BANDED DIORITIC GNEISS: SOME MASSIVE PATCHES	U-PB SPHENE	1865
F ₃		SE	ASYMMETRICAL FOLDS: ANTI-CLOCKWISE ROTATIONAL SENSE		DISCORDANT QUARTZOFELDSPATHIC VEINS	U-PB SPHENE	1880
F ₄		N-E	ASYMMETRICAL FOLDS: CLOCKWISE ROTATIONAL SENSE		DISCORDANT QUARTZOFELDSPATHIC VEINS		
F ₅		NNE	OPEN FOLDS ASSOCIATED WITH INCIPENT CONJUGATE SHEAR		CONCORDANT TO DISCORDANT QUARTZOFELDSPATHIC VEINS AND APROLOGRANITE		
F ₆			ANGULAR FOLDS (ASYMMETRICAL)		DISCORDANT, COARSE GRAINED BIOTITE, K-FELDSPAR PEGMATITE	U-PB MONAZITE U-PB ZIRCON AND MONAZITE	1870 1840
F ₇		N	OPEN WARPS		INTRUSIVE APLITE VEINS PERVASIVE AND INTRUSIVE APROLOGRANITE	RB-SR MUSCOVITE RB-SR	1810 1800
					DISCORDANT, MUSCOVITE-BEARING, K-FELDSPAR PEGMATITE	WHOLE-ROCK	1600

FIGURE 2. Fold sets, geometry, axial trend and related neosomes with their ages; Svecokarelian migmatites, S Finnish archipelago. (After Hopgood et al., 1983.)

accompanying igneous emplacement and agmatization.

The neosomes emplaced at different times, which constitute an integral component of migmatites, may be distinguished one from the other by reference to their relationships to the structures in the succession. These relationships are often very clearly shown, particularly in agmatites (Hopgood and Bowes, 1978). For example, the position of a particular neosome in the deformational sequence can be determined by its discordance to particular structures in the palaeosome or to an earlier-formed neosome which, for example, postdates the earliest structures in the palaeosome yet was itself deformed before the emplacement of the later neosome. In such cases neosomes that are apparently lithologically the same can be distinguished. Neosome-forming events throughout the developmental history of migmatites can thus be placed in relative order and their petrogenesis considered in relation to their isotopic age (Hopgood et al., 1983).

The fold patterns in migmatites represent the cumulative effects of a succession of structures. In southern Finland as many as 15 sets of structures and 10 phases of neosome emplacement have been identified (Hopgood, 1984) and it has been possible to correlate the fold and neosome sequence throughout a considerable region where the exposure is excellent. Seven of the most prominent of these fold sets (here labeled F₁ to F₇) are given as examples of the kinds of structures developed in migmatites (Fig. 2). Although they represent neither the

complete succession, nor necessarily include either the first or last sets of structures, they are nevertheless representative of fold sets typical of structural successions in the migmatites of southern Finland and show some general correspondence with fold sets in other migmatite complexes. They range from intrafolial, isoclinal, asymmetrical, angular folds to open warps. The type of deformation shows broadly sequential transition from ductile to brittle, and the change from early pervasive neosome to the later more discrete neosome is a reflection of changing *P-T* conditions with crustal depth. The latest structures are shears and fractures.

F₁ structures. These are typically isoclinal and nearly always intrafolial (Fig. 3). They can often be subdivided into at least two groups: the earlier are usually extremely attenuated and are sometimes distinguishable from the latter by virtue of having consistently different axial attitudes. The earlier folds normally have slender profiles and may have sharply tapered hinges, with ratios of hinge to limb thickness of between 10:1 and 40:1. They often appear solely as rootless hinges or complete closures in the hinge and limbs of the later folds and may also differ in showing obvious asymmetry and/or rounder hinges.

F₂ structures. It is usual to find the early attenuated isoclinal folds folded by tight, sometimes isoclinal, symmetrical and asymmetrical upright folds with interlimb angles of 10–20° (Fig. 3). Their axial planes are generally parallel to the dominant foliation. The fold hinges may have generally round, even lobate,



FIGURE 3. Early isoclinal folds (F_1) folded by F_2 ; Gråkløkken, S Finland.

profiles and often can be seen to be compound structures. This is the result of modification of folds that probably originated flexurally but have been affected by passive deformation (flow and slip more or less parallel to their axial planes) later in their history. The fold profiles have also been modified less subtly by dislocation following movement on discrete, widely-separated surfaces that are subparallel to the existing fold axial planes and associated with quartzofeldspathic veins. Mesoscopic folds of this set range from a centimeter or two, to 2–3 m in amplitude. In the least-modified of these, the ratio of hinge to limb thickness is between 1.5:1 and 2:1. Characteristically, folds of this set are associated with gray leucocratic neosome that appears to have developed penecontemporaneously. The foliation in the gray gneiss so formed generally is parallel to both limbs and hinges of F_2 except where modification by later deformation has been strong, in which case the foliation can be axial planar to the folds.

F_3 structures. The trend of F_2 axial traces is deflected, along with the foliation, by asymmetrical S-folds (Fig. 4) with curved axial planar traces and steeply-plunging axes. Folds range

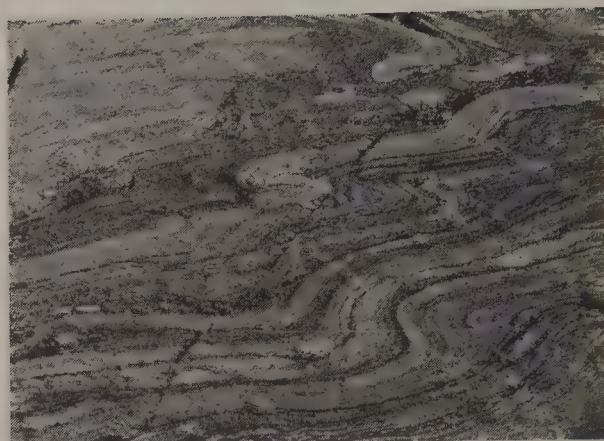


FIGURE 4. Asymmetrical (F_3) folds; Björkholmen, S Finland.

between close and open (20–90° interlimb angles) and have a hinge-to-limb thickness ratio of about 1:2. The profiles of F_3 vary from being almost boxlike in form to flattened, nearly isoclinal with axial planes drawn into near-parallelism with the dominant foliation during later deformation. The folds may be accompanied by a faint axial planar foliation and mineral growth that cuts across earlier-formed fabric, and the hinge zones of some are partially obliterated by irregular patches of leucosome.

F_4 structures. Steeply-plunging asymmetrical folds, demonstrated to be later than F_3 folds on the basis that they refolded them, have axial planar traces trending approximately 45° to the dominant foliation. They are close, Z-folds (Fig. 6) with interlimb angles of about 50° and their profiles tend to be more angular than those of F_3 . The difference between hinge and limb thickness is less pronounced than in F_3 folds, and F_4 axial planar traces are more nearly rectilinear than those of F_3 and commonly associated with quartzofeldspathic veins. Asymmetrical folds deforming early tight structures are common features of fold sequences in other migmatite terranes also (Fig. 5).

F_5 structures. Open symmetrical folds with an interlimb angle of about 80° and an axial



FIGURE 5. Early folds (F_1 , F_2 , F_3) in migmatites of the Lewisian complex; Harris, Scotland.

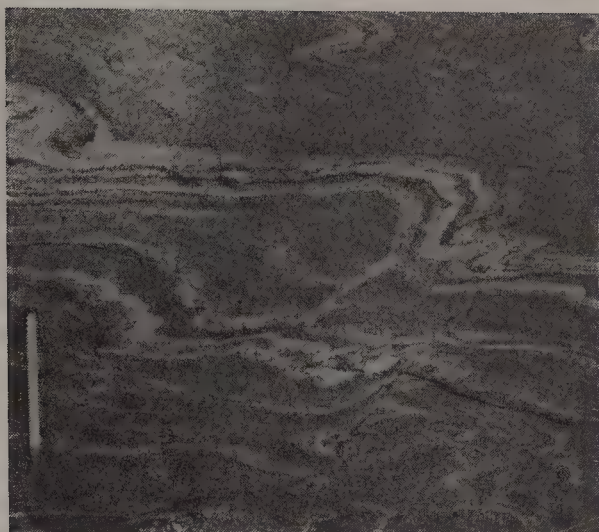


FIGURE 6. Asymmetrical (F_4) folds; Ängholmen, S Finland.

planar trace trending almost 90° to the dominant foliation are superimposed on the structures previously described. They have a wavelength-to-amplitude ratio of about 6:1 and show a distinctive blurring of the foliation on one or both limbs caused by incipient shear and slight development of potassic neosome (Fig. 7). This leucosome locally forms irregular ramifying pink blotches, pods, and discordant veins that trend parallel to the fold limbs and coalesce to form masses of equant porphyroblastic matrix. It is normally easily distinguishable in terms of texture and color from the earlier syn- F_2 gray



FIGURE 7. Small, open F_5 fold showing incipient development of neosome (arrowed) along the trace of one fold limb; Langören, S Finland.

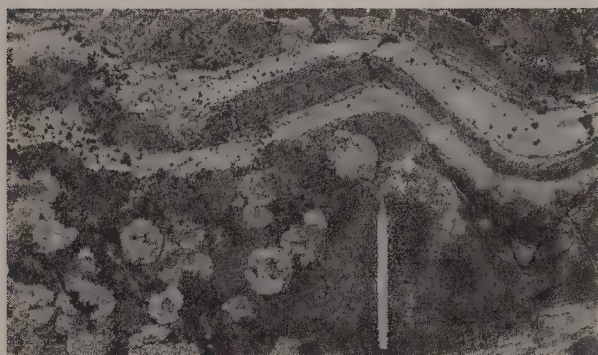


FIGURE 8. Angular (F_6) fold typical of structures late in a structural succession; Skepparnäs, S Finland.

tonalitic neosome, especially where the two are associated in polyphase agmatites (Hopgood and Bowes, 1978).

F_6 and F_7 structures. Towards the end of the deformational sequence, a set of open folds was formed with angular or slightly round hinges (Fig. 8) and interlimb angles of 90 – 110° and a wavelength-to-amplitude ratio of approximately 2:1. They are nearly planar-limbed, one limb or the other being associated with incipient dextral or sinistral shear, and their axial planar traces trend consistently 20° anticlockwise from those of F_5 .

These and earlier folds have all been subjected to reorientation on the limbs of gentle warps (F_7) that have axial traces which deflect the foliation through angles of up to 20° over a distance of about 10 m. F_7 folds have a wavelength-to-amplitude ratio of approximately 20:1 and appear to be small-scale examples of the large, open structures (wavelengths of kilometers) dominating the structure of SW Finland, and are typical of structures developed in other migmatite terranes (Fig. 9).

Subvertical, SSE-trending ductile shears with which are associated tight flow folds succeed F_7 and these in turn are offset by a series of steeply



FIGURE 9. Early and mid-succession folds with gently inclined axial planes folded by later upright open folds and warps (comparable to F_7 in the Finnish succession) with almost vertical axial planes; Archaean Preketilidian migmatites, SW Greenland.

dipping brittle fractures and late conjugate shears.

Assessment. While the structures described include only some of the more distinctive types of folds, they nevertheless serve to illustrate the styles of the components of a structural pattern typical of Precambrian migmatite and gneissic terranes seen throughout the world (e.g., Hopgood, 1970; Hopgood and Bowes, 1972; Gower and Clifford, 1981; Halden et al., 1982; Suo et al., 1982) and illustrate the basis for differentiating between neosomes emplaced at different times during the developmental history of a migmatite complex.

Significance in elucidation of crustal evolution

Determination of the relative chronology of successively formed structures forms the basis for determination of the relative chronology of metamorphism and episodes of igneous emplacement. This in turn provides a basis for isotopic studies to establish time spans of tectonothermal episodes and intervals between phases within episodes (e.g., Hopgood et al., 1983). Erection of a structural chronology also forms the basis for demonstrating changing *P-T*-depth conditions within one part of an orogen and from place to place within an orogen. In addition, the nature of structures developed can provide evidence relating to the movement of lithospheric plates. For example, the evolution of the migmatite complex of southern Finland took place in response to alternating compression and slip (Fig. 2), suggesting a combination of plate collision and strike-slip fault movement (Hopgood, 1984). Integration of the neosome emplacement with this structural evolution provides a basis for tracing the changing pattern in the nature and depth of magma source and so of helping to elucidate the overall development of major orogenic regimes.

A. M. HOPGOOD

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Cross-references: *Anatexis*; *Migmatite*; *Migmatite*—origin of neosome.

MINERAL THERMOMETRY—See Vol. IVA

MNEMONICS, ACRONYMS

Biopyribole—a term coined by A. Johannsen in 1911 in his classification of igneous rocks to indicate the presence of *biotite*, *pyroxene* and/or *amphibole*; now used in a mineralogical sense (see **Biopyribole**, Vol. IVB).

Cafemic—refers to an igneous rock or magma that contains Ca, Fe, and Mg.

Dofemic, *dosalic*, *perfemic*, *persalic*, and *salfemic*—used to designate classes and *dosemic*, *dohyaline*, *domoikic*, *dopatic*, *doxenic*, and *sempatic* to designate variants in the CIPW classification of igneous rocks (see **Petrochemical calculations**).

Felsic—refers to rocks essentially composed of *feldspar*, *feldspathoid* and *silica*; it is not synonymous with *salic* which refers to normative compounds calculated from a chemical analysis.

Femic—derived from *ferromagnesian*, and refers to an igneous rock in which Fe- and Mg-rich minerals (*pyroxene*, *olivine*) are the

major components of the norm; the corresponding term for ferromagnesian minerals in a rock is *mafic*.

Kamafugite—a collective name derived from *katungite-mafurite-ugandite*.

KREEP—nonmare lunar basalt characterized by unusually high proportions of potassium (K), rare-earth elements (REE) and phosphorus (P) (see **Lunar petrology**).

Mafic—refers to rocks composed of one or more ferromagnesian, dark-colored minerals; it is not synonymous with *femic* which refers to normative compounds calculated from a chemical analysis.

Pyribole—similar to biopyribole but without biotite.

Pyribole, *pyriclasite*—pyroxene-hornblende (amphibole)-plagioclase and pyroxene-plagioclase rocks that are high-grade metabasites, the end members containing >2/3 of the particular component (see **Granulite**).

Pyrolite—a term for the modal composition of the upper mantle derived from pyroxene and olivine (see **Pyrolite**).

Salic—a term derived from silicon and aluminum used to refer to certain light-colored Si- or Al-rich minerals present in the norm of igneous rocks (e.g., quartz, feldspars, feldspathoids) and to rocks having one or more of these minerals as major components of the norm.

Sial (sialic)—a term derived from silicon and aluminum used to refer to the upper layer of the continental crust.

Sima—a term derived from silicon and magnesium used to refer to the lower portion of the continental crust or the oceanic crust; now largely obsolete, as are *sialma* (*sialsima*), which referred to those parts of the Earth's crust intermediate between sial and sima, and *sifema* (silicon, iron (Fe), magnesium), which referred to an ultrabasic layer.

D. R. BOWES

MOHOROVIČIĆ DISCONTINUITY (MOHO)

The Mohorovičić discontinuity is the boundary (also referred to as the M-discontinuity, or Moho) that separates the Earth's crust from the subjacent mantle at depths of 30–50 km below the continents and 5–10 km below sea level in the oceans. Above it, measured velocities of compressional waves are usually <ca. 7 km/sec, while below it they are ca. 8 km/sec. On the basis of ophiolites representing samples of the ocean crust, the “petrological Moho” corresponds with the base of layer 3 (gabbro and

various ultramafic differentiates and cumulates) and the top of layer 4 (mantle peridotites). The petrologically poorly-defined “geophysical Moho” corresponds to the disappearance of feldspar in the downward gradation of gabbroic cumulates to ultramafic cumulates in layer 3. Estimates of the thickness of the boundary zone differ from a few hundred meters to a kilometer or even several kilometers. It was named in honor of its discoverer (A. Mohorovičić) who noticed, when examining the earthquake records of the Kulpa Valley Earthquake of 8th October 1909 in Yugoslavia, that the velocities of seismic compressional waves increased abruptly at a certain depth within the Earth. This seismic discontinuity has since been recognized almost everywhere geophysicists have investigated the Earth's structure.

D. R. BOWES

Cross-references: *Mid-ocean ridge and ocean-floor petrology*; *Oceanic crust*; *Ophiolite*; *Upper mantle—experimental petrology*.

MONZONITE (SYENODIORITE)

Characteristics

Monzonites are uncommon intermediate intrusive igneous rocks with quartz <5% and alkali feldspar: plagioclase between 35 and 65%. In bulk composition they may be indistinguishable from some two- and one-feldspar syenites, a distinct shortcoming of the Streckeisen (1976) classification (see **Syenite**). With increasing quartz, monzonite grades into quartz monzonite and granite, and with increasing plagioclase into *monzodiorite* and *diorite*. The term syenodiorite, although strictly synonymous, tends to be used for rocks in alkaline suites (i.e., in association with syenites and syenogabbros) while monzonite tends to be applied in calc-alkaline assemblages (i.e., in association with granodiorites). Feldspathoid-bearing syenodiorites also occur. The name monzonite is due to C. W. von Buch (see Johannsen, 1937) after Monzoni, Tyrol.

In American usage, quartz monzonite corresponds to granite of Streckeisen (1976) and is almost synonymous with adamellite of European geologists. It does not correspond to quartz monzonite in the IUGS nomenclature.

Calc-alkaline monzonites have similar mafic mineralogy (hornblende and biotite) to the granodiorites and granites, but less quartz. These rocks are quartz-poor, two-feldspar assemblages in which the discrete plagioclase phase may be andesine to albite together with perthitic orthoclase. Alkaline equivalents

(syenodiorites) may contain plagioclase rimmed by ternary cryptoperthitic anorthoclase feldspar, or set with interstitial alkali feldspar. Such rocks may show their alkaline tendencies by the presence of Fe-rich biotite, ferroaugite, and Mg-Fe olivines. Spene and particularly apatite are common accessory minerals.

Distribution and genesis

True quartz-free calc-alkaline monzonites are uncommon rocks that may occur as discrete plutons but more commonly as local or marginal variants to more acidic plutons. Quartz monzonites (of American usage—elsewhere termed adamellites) are, however, important members of the assemblages of tonalite and granodiorite that characterize orogenic belts (such as the Andes—Cobbing and Pitcher, 1972), and for which an origin by deep crustal anatexis is favored by many authors (see **Diorite and tonalite**).

Alkaline syenodiorites are always of restricted extent, even in areas where syenitic rocks predominate and occur together with potential gabbroic parents (Upton, 1974). A few examples of continuous gradations from gabbro to syenite via syenodiorite occur in the Gardar province, S Greenland. Such rocks lie on a trend towards larvikitic syenites, into which they may grade.

Other types

Amherstite—a monzonite with andesine and antiperthite (Amherst Co., Virginia, U.S.A.).

Kjelsasite—a monzonite with affinities to larvikite.

Masanite—a quartz monzonite with phenocrysts of zoned plagioclase and corroded quartz (Ma-san-po, Korea); the variety *masanophyre* has phenocrysts of oligoclase rimmed by orthoclase in a groundmass of hornblende and spene.

Sörkedalite—a troctolitic monzonite or olivine monzodiorite rich in apatite and Fe-Ti-oxides.

Ukrainite—a monzonite with <20% quartz.

Vallevarite—a leucocratic monzonite composed mainly of andesine, microcline, and antiperthite with small proportions of diopside, biotite, and apatite.

Windsorite—a leucocratic monzonite with a minor proportion of biotite (Windsor, Vermont, U.S.A.).

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Cross-references: *Alkaline rocks—undersaturated; Diorite and tonalite; Intermediate igneous rocks; Plutonic rocks—classification and nomenclature; Syenite.*

N

NEPHELINE

A *nephelinite* is a hypabyssal or extrusive strongly undersaturated igneous rock composed essentially of nepheline and pyroxene. If olivine is present, the rock may be called a nepheline basalt, but the term olivine nephelinite is preferable; it is also preferable to *ankaratrite*, a term used for a biotite-bearing melanephelinite (Ankaratra, Malagasy). Many nephelinites are melanocratic rocks (melanephelinites, which have the plutonic equivalent, ijolite, grading to melteigite). With the presence of plagioclase they grade into nepheline basanites (with olivine) and nepheline tephrites (without olivine)—*atlantite* is a dark-colored nepheline basanite with a preponderance of dark-colored minerals, while *lualaba* is a tephritic leucite nephelinite with >50% mafic minerals. Alkali feldspar-bearing types are called *phonolitic nephelinites*. Many nephelinites are agglomeratic.

Mineralogy

Nephelinites are usually holocrystalline, although some glass may be present. Essential nepheline may be allotriomorphic and interstitial or present as euhedral grains, sometimes enclosed by mafic phases. The most important mafic phase is a diopsidic pyroxene, often with slightly acmitic margins. Biotite (sometimes poikilitic) or corroded hornblende may be present. In olivine nephelinites, well-formed olivines, with alteration to serpentine and ore, may occur. Primary intergrown magnetite-ilmenite crystals also appear. Melilite, leucite, and sodalite-group minerals occur frequently, and the latter may be abundant. Such rocks have been called hauyne or nosean basalts. Eudialyte is also listed by Johannsen (1938). Alkali feldspar, where present, is interstitial to other phases. Minor calcic plagioclase may also be present.

Distribution and genesis

The greatest development of nephelinitic rocks is in the African rift system, where there are numerous nephelinitic volcanoes, associated with carbonatite-ijolite plutonic complexes.

Common associated extrusive rocks are nepheline and melilite basalts, and melilitite. This association is rare in the oceanic regions, and when it does occur the volume of nephelinite and its associates is trivial compared with normal basalt. The extreme of undersaturation shown by nephelinite represents a problem for schemes of fractional crystallization from basaltic rocks (discussed by Bailey and Schairer, 1966) and the field setting is consistent with deep partial melting in the mantle. Green (1970) shows that nephelinite may be formed by partial melting of a hydrous model mantle (pyrolite) at pressures in the range 18–35 kb. The mineralogy of mantle xenoliths from rift valley volcanic rocks, and the trace-element chemistry of nephelinites, suggest that they have their source in mantle “enriched” in incompatible and characteristic trace elements, perhaps as a result of migration through the mantle of CO₂-rich fluids, which in turn give rise to the very commonly associated carbonatite volcanism (Le Bas, 1977, 1987; Bailey et al., 1980).

Other terms

Grazinite—a mesocratic phonolitic nephelinite containing analcime but not olivine.

Leucitonephelinite—a nephelinite containing leucite.

Nemafite—a volcanic rock composed of nepheline and mafic minerals (i.e., a nephelinite).

Nephlinolith—an effusive rock composed entirely of nepheline.

Ngurumanite—a nepheline-pyroxene rock with a mesostasis of serpentine and chlorite and vugs of calcite and analcime; a hypabyssal ankaratrite (or melteigite).

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Cross-references: *Alkaline rocks—undersaturated; Basalt; Carbonatite; Feldspathoidal rocks—genesis; Ocean-island igneous rocks; Phonolite; Rift valley volcanism; Tephrite, basanite.*

NEUTRALITY—pH AND TEMPERATURE VARIATION

Almost all minerals (silicates, carbonates, sulfides, oxides) are salts of weak acids and relatively strong bases. Thus, their solutions in pure water at room temperature will be alkaline because of hydrolysis reactions, and their solubilities are a function of hydrogen ion activity. To understand quantitatively the physical chemistry of such solutions, it is necessary to know the values of the ionization constant of water, which dissociates according to the equation



or



the latter reaction indicating that the aqueous proton is always hydrated. Only the first dissociation step of water is of quantitative significance and the equilibrium constant describing this reaction is

$$[\text{H}^+][\text{OH}^-] = K_w$$

where the terms in brackets represent the activities or thermodynamic concentrations of the species.

At 25°C in liquid water, $K_w = 1.008 \times 10^{-14}$ and is usually quoted as 10^{-14} . From this, it follows that the hydrogen ion activity of pure water is 10^{-7} mol/litre and solutions that have a hydrogen ion concentration equal to that of pure water are termed *neutral*. Neutral solutions are produced by the salts of strong acids and strong bases (NaCl, KCl, etc.). Salts of weak acids and strong bases (e.g., CaCO_3) give alkaline solutions, and salts of strong acids and weak bases (e.g., NH_4Cl) produce acid solutions.

In discussing hydrogen ion concentrations or activities, the quantity pH or $-\log_{10}[\text{H}^+]$ is often used. Thus the pH of pure water at 25°C is

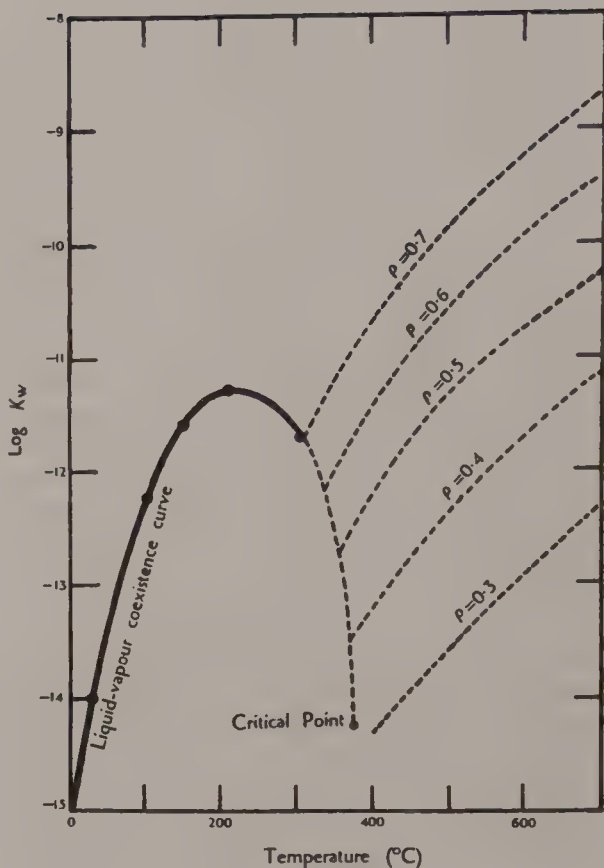


FIGURE 1. Ionic dissociation of water at high temperatures; solid lines—experimental data of Noyes (1907); dashed lines—calculated by Franck (1961).

7 and this represents neutrality. This definition of pH in terms of hydrogen ion activity presents some difficulties, in that pH is a theoretical concept and it is impossible to measure the activity of a single ion in solution. But very reasonable practical devices are used to overcome this problem.

If neutrality is defined as above, $\text{pH} = 7$ represents neutrality only at 25°C and 1 atm pressure. Under any other conditions K_w must be known. Some data are available over the geological range, in part measured and in part estimated by theoretical methods (Fig. 1). If P - V - T relations for water are known, then at 600°C and a density of 0.7, pressure ca. 3,300 bars, $K_w = 6 \times 10^{-10}$ and the pH of a neutral solution will be 4.61.

The studies of Franck (1961) have shown that salts behave quite differently at high temperatures and moderate pressures. Most salts like NaCl are present in solutions mostly in molecular forms. Strong acids and bases such as HCl and KOH are weak and also present mainly as neutral molecules. Thus, in general, there is a tendency towards molecular behavior in aqueous solutions at high temperatures.

W. S. FYFE

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NOSEANITE

A *noseanite* is a feldspar-free and olivine-free basalt that contains abundant nosean. *Noseanolith* is the term used for an effusive rock composed almost entirely of nosean. A *noselitite* is an effusive rock mainly composed of nosean and a pyroxene, sometimes with amphibole present. These rocks are part of an association of strongly undersaturated rocks in which haüyne, leucite, melilite, nepheline, nosean, and sodalite (and cancrinite) variably express the inadequate silica content in the rock for the complex formation of alkali feldspar.

D. R. BOWES

Cross-references: *Alkaline rocks—undersaturated; Feldspathoidal rocks—genesis; Haüynite; Leucitite; Melilitite; Nephelinite; Sodalitite.*

NUCLEATION OF METAMORPHIC MINERALS

Nucleation is the term referring to the sequence of interlinked physical and/or chemical changes that together result in the formation of a new particle (a nucleus), capable of subsequent growth to form a crystal.

It is not possible to make a direct experimental investigation of the nucleation processes that occur in most of the system of greatest geological interest. Discussion of the significance of this process during reactions in natural systems must be based, therefore, on indirect evidence or through analogy with much simpler chemical systems. It is important to realize the limitations inherent in such comparisons.

Since nucleation is the process that initiates product formation, it exerts some control upon reaction rate and is, therefore, considered by the methods of chemical kinetics. If nucleation is a relatively facile process, the product formed

may consist of a large number of small crystallites, whereas the same total quantity of product arising from a reaction in which nucleation occurs less readily may consist of a smaller number of relatively large crystals. Kinetic factors may thus influence the size distribution of product particles within the rock.

From standard theory of kinetics, the rate of nucleus formation (dN/dt) may be expressed:

$$\frac{dN}{dt} = AC_x f(a_1, a_2, \dots) e^{-E/RT}$$

where the preexponential A includes due allowance for the number of nucleation sites per unit volume, a vibration frequency in the reaction coordinate, steric restrictions, entropy, etc.; C_x is the availability of reactant species (perhaps an appropriate configuration at crystal surfaces, or a group of atoms or ions); $f(a_1, a_2, \dots)$ is a measure of availability, i.e., concentration, of various other possible participants in the rate-limiting step; $e^{-E/RT}$ is the fraction of reactant situations in which the energy barrier (E) to reaction is surmounted.

The meaningful application of this expression to quantitative geologic studies is not straightforward. For example, the term "concentration" in any heterogeneous system requires careful definition. During metamorphic reactions, reactant may be transported by random or semirestricted migration through the substrate by intergranular paths or diffusion in the defect structure of the minerals. The ease of movement may, therefore, be an important factor in the availability of reactant and thus in the effective concentration during reaction. Even in pure solids, the kinetics of nucleation vary from one system to another, (Brown et al., 1980); it is probable that further complexities arise in heterogeneous systems. Moreover, not every nucleus develops to form a crystal; some may be ingested during the period of initial instability. This is because the small group of atoms formed initially possesses a relatively large surface-to-volume ratio and a certain critical size must be reached before the developing particle achieves stability. This critical radius is small for silicates (for the transition of cristobalite to quartz at 1000 K, it has been estimated as 15 nm but it is considered that the radius decreases with increasing supersaturation (Rast, 1965). However, the overall importance of concentration is indicated by such phenomena as the well-known tendency for garnets rich in Mn to form relatively large numbers of crystals per unit volume (Spry, 1969). This may be because the large Mn ion is particularly stable in the garnet lattice and thus a high Mn concentration produces more stable

nuclei with an increased chance of developing as product crystals. Concentration is also involved with nucleation in a more subtle way. It is usually accepted that the instability of the initial nucleus imposes an energy barrier or other restriction upon development, so that subsequent growth of an existing nucleus occurs more readily than a further nucleation step. Such growth must then reduce the availability of the elements required for nucleation in the immediate vicinity of the growing crystal. If transport is sufficiently rapid, the probability of a new nucleation process is also reduced. It is possible, therefore, that rapid heating of a rock would result in the development of a relatively large number of new nuclei, if other factors remained constant. At lower temperatures, the longer time intervals available would allow material from a larger domain to be incorporated within a smaller number of nuclei. Such hypotheses should be capable of testing (Jones et al., 1972).

The formation of a nucleus capable of growth is thus a statistically improbable event and depends on such factors as the availability of a suitable site and favorable environmental conditions such as temperature, pressure, and adequate availability (including concentration) and diffusion rate of reactant species. Nucleation probably occurs preferentially at (1) sites of particular reactivity or instability within or at surfaces or interfaces of preexisting crystals (cf., Heard, 1963) and/or, (2) by epitaxial growth at a region of crystal surface that approximates to the lattice structure and/or composition of the nucleated phase (cf., Frondel, 1940).

It is generally accepted that dislocations, particularly at the point of emergence on surfaces, are favored sites for nucleation, owing to local strain fields and lattice distortion at which the crystallographic requirements of the lattice of the nucleus phase can be relatively easily satisfied (Brown et al., 1980). This is likely to be particularly important for those reactions in which the new phase has a different volume from that of the precursor crystal and thus produces a local region of elastic strain.

The importance of interfaces to nucleation has been discussed by McLean (1965), who showed that nucleation rate depends on the cube of the interfacial energy and the square of the free energy change. He also pointed out the definite relationship between γ (interfacial energy/unit area between new phase and matrix) and ΔF (free energy liberated by the formation of unit quantity of new phase). It can thus be deduced that under metamorphic conditions, for which most, if not all, reactions involve a small overall change in free energy, nucleation can only occur where the interfacial

energy is low, perhaps within the range of $0.005\text{--}0.1\text{ J/m}^2$; this requires a good matching of interfaces. It also follows that crystals could form in rapidly cooled magmas with a much higher interfacial energy, since the free energy change is correspondingly higher. If a phase tends to nucleate at a particular type of interface, the distribution of this phase will be at least partly controlled by the distribution of that specific preferred interface, i.e., the microstructure and mineral composition of the rock.

K. A. JONES
A. K. GALWEY

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Cross-references: *Metamorphic reactions; Metamorphism; Phase transformation.*

NUCLEATION OF MINERALS IN MAGMAS

When a totally liquid magma is cooled it passes through a particular temperature at which the magma is saturated with respect to a dissolved mineral phase. However, crystallization of the mineral does not occur until there has been further cooling to a level of supersaturation and a temperature at which small, stable nuclei of crystals begin to form and

grow, i.e., the process of *nucleation*. Because the crystals nucleate at a temperature below that at which they would melt in the magma, the liquid is said to be “supercooled” and so is in a metastable state. Metastable supercooled magma exists because at low levels of supersaturation any crystal nucleus that may form is soon disrupted by molecular motion before it can grow to a stable size. Supersaturation is the driving force for crystal growth and, within certain limits, the greater the degree of supersaturation the faster is the growth rate. Hence, there is a level of supersaturation at which there is a greater probability of crystal nuclei growing than of them being disrupted over a period of time. The extent of the metastable region is dependent on several factors, including the rate of cooling and the movement of the magma.

In addition to this process of spontaneous nucleation of a crystal at some degree of supersaturation (homogeneous nucleation), nucleation may also be initiated at a degree of supersaturation less than that required for spontaneous nucleation, by the presence of “seed” nuclei or by the liquid receiving some sudden movement or shock (heterogeneous nucleation).

Theory

Nucleation of crystals may commence if the production of crystals lowers the free energy (G) of the system. However, the initial step of forming a crystal nucleus requires an increase in the free energy of the system relative to the value at equilibrium, because of the added surface free energy of the solid–liquid interface. Hence, crystallization will proceed if the reduction in the free energy of the whole system is greater than the excess free energy created by the production of the solid–liquid interface, and such a condition exists only when the melt or magma is sufficiently supersaturated with respect to one (or more) mineral phase(s). The free energy change (ΔG) of the system for the formation of a crystal of the same composition as the melt can be given (for a spherical nucleus) by

$$\Delta G = 4\pi r^2 \sigma + \frac{4}{3}\pi r^3 \Delta G_v$$

where r is the radius of the crystal particle, σ is the surface free energy per unit area, and ΔG_v is the free energy change of transformation per unit volume. The equation shows, for a supersaturated solution (ΔG_v negative) and for small values of r , that the free energy change increases with increase in r until some critical radius (r^*) is reached, after which the free energy decreases with further increase in r (Fig.

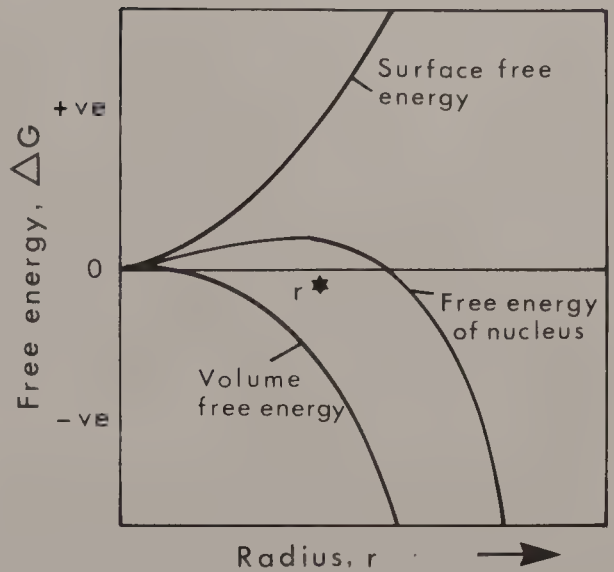


FIGURE 1. Relationship of free energy and radius of nucleus; r^* is the critical radius.

1). In heterogeneous nucleation, the interfacial energy between the seed crystal and liquid usually acts to lower the free energy of the system and cause nucleation at a lower level of supersaturation.

Kirkpatrick (1981) gives a review of the theory of nucleation, but as yet it has not proved useful as a predictive tool for petrologists or mineralogists (Maaløe, 1975).

Experimental work

The rate of nucleation as a function of supersaturation has been shown to reach a maximum, followed by a decline with increasing supersaturation or undercooling. An example of this relationship has been given by Fenn (1977), who studied the variation in nucleation rate (or more strictly the nucleation density (N), i.e., the number of nuclei created per unit volume over the duration of an experimental nucleation–growth step) with degree of undercooling (ΔT) (Fig. 2). His experiments involved

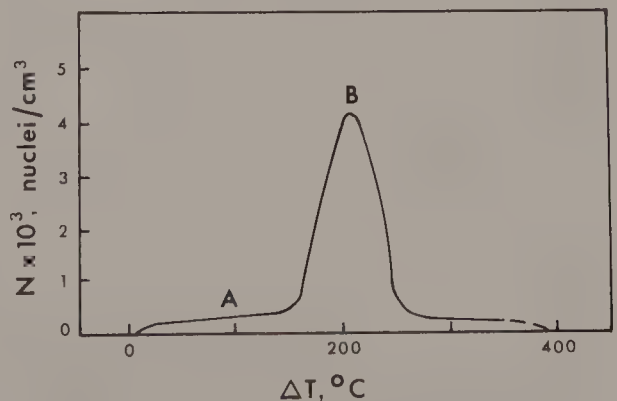


FIGURE 2. An example of the variation in nucleation density (N) with degree of supercooling (ΔT), for alkali feldspars in a hydrous Na–K–Al silicate melt.

the nucleation of alkali feldspars in synthetic, hydrous Na-K-Al silicate melts and showed that the maximum in the nucleation-density curve shifts towards the liquidus temperature with increase in water content of the melt. The decline in the density at higher degrees of undercooling (Fig. 2) may be explained by the greater viscosity of the melt with large supersaturations, which in turn rapidly increase the activation energy for molecular motion across the incipient nucleus-liquid interface. Experimental work has also shown that nucleation is dependent on several other factors.

Crystal structure. Orthosilicates (e.g., olivine) tend to nucleate more readily than chain silicates (e.g., pyroxenes) and these more readily than framework silicates (e.g., feldspars), in synthetic and natural silicate melts. Differences in crystal structure can lead to the metastable nucleation of a nonequilibrium phase and suppression of an equilibrium one, e.g., metastable olivine occurs in the groundmass of some mid-ocean ridge basalts but the equilibrium orthopyroxene or pigeonite phase is absent.

Melt structure and composition. There is some evidence to suggest that the nucleation rate decreases as both the silica content and the degree of polymerization in the melt, increase. Water content also affects nucleation.

Cooling rate. It is clear that the cooling rate of a melt will have a significant effect on a time-dependent process such as nucleation. Very rapid cooling can lead to production of a glass with no crystal nuclei. Slow cooling will tend to allow the nucleation and growth of equilibrium phases.

Crystal nucleation will not necessarily occur immediately when a magma is cooled rapidly in one step from above the liquidus to a temperature at which nucleation is possible. There is a delay in nucleation at the new temperature. Donaldson (1979), for example, has shown that the delay of olivine nucleation in basaltic melts increases systematically with decreasing degrees of supercooling, cooling rate, and olivine content of the melt, as well with decrease in the difference between the initial temperature and the liquidus temperature.

Applications

In their discussion of the origin of textures in the rocks of the Skaergaard intrusion, E Greenland, Wager and Brown (1968) related some textures to the varying rate of nucleation with degree of supersaturation of a mineral phase. There the chilled marginal gabbro has

large numbers of small crystals of plagioclase feldspar and olivine, poikilitically enclosed by relatively few crystals of pyroxene and ilmenite. The intruding magma was rapidly chilled against the country rock and became supersaturated with olivine and plagioclase such that many nuclei of these minerals formed and then grew (region B of Fig. 2). The magma did not become saturated with respect to pyroxene and ilmenite until further cooling had taken place and by then the country rock had warmed up and the temperature was falling more slowly. These conditions give rise to a low nucleation rate (region A of Fig. 2) and the slow cooling allowed the few nuclei of pyroxene and ilmenite to develop into large poikilitic crystals. Textures in other igneous rocks, especially in basalts from the ocean basins and from the Moon, have been interpreted using nucleation sequences and variations in growth rates of the constituent minerals.

PAUL HENDERSON

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Cross-references: *Magma*; *Skaergaard intrusion*; *Textures of igneous rocks*.

NUÉE ARDENTE

The term *nuée ardente* was first introduced by Lacroix in 1903 (and later amplified in the fuller work of 1904) to describe the mixtures of gases and pyroclastic material that flowed down the flanks of Mt. Pelée in its volcanic activity of 1902-3. Its aptness has been questioned, since the "glowing cloud" of gases rising into the atmosphere and reflecting light from the incandescent material at its base is much less important than the suspension of magmatic (and solid) fragments in hot gases that flows down the mountain slopes (Fig. 1). The term is,

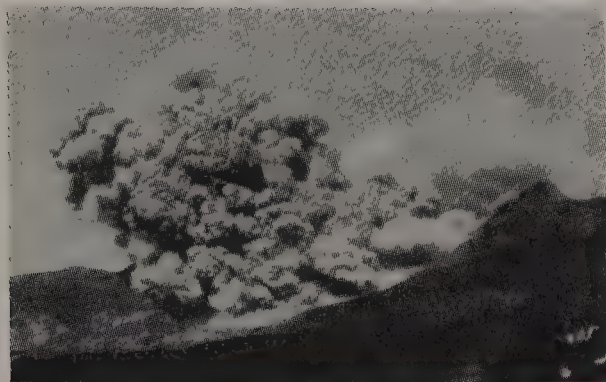


FIGURE 1. A small typical *nuée ardente* showing a cloud of gases rising from a flow; Mt. Pelée, Martinique, West Indies, 8th February, 1930.

however, well established and, despite some initial misunderstanding, it is now universally appreciated that it has a dual significance, applying to the flow of pyroclastic material as well as the cloud of gases that rises from the flow.

The pyroclastic flow in the lower portion of a *nuée ardente* frequently moves at a high velocity. The combination of high temperature, high velocity, and the weight of magmatic material moving in the flow often makes this type of volcanic activity both dangerous and destructive. It is capable of grooving or gullyng rock surfaces across which it moves as well as demolishing trees, buildings, etc., in its path. When the flow comes to rest it does so with a virtually flat featureless upper surface, e.g., the Valley of Ten Thousand Smokes following the 1912 eruption of Novarupta (Katmai) in Alaska.

The interest of the geologic world was greatly aroused by this apparently novel volcanic phenomenon in the West Indies in 1902, especially on account of the sudden destruction of the town of St. Pierre (Martinique) on 8th May with the death of some 30,000 people (as well as the similar type of eruption from La Soufrière on nearby St. Vincent Island killing 1,500 the previous day). Shortly after the St. Pierre disaster T. Anderson and J. S. Flett were sent to the area by the Royal Society of London, and their description of a *nuée ardente* witnessed on 9th July 1902 represents one of the first accurate accounts of this feature (Anderson and Flett, 1902).

This type of volcanic eruption (later called the *peléan* type) had been noted before 1902 but had not received much attention. Rittmann (1962, p. 29) observed that Sir William Hamilton described a mobile suspension of solid particles in hot gases at Vesuvius in 1773. Fenner noted that F. von Wolf's description of the Cotopaxi eruption was an excellent account

of a major eruption of *Peléan* type. These early accounts had, however, made little impact prior to the eruption of Mt. Pelée (and La Soufrière) in 1902. Further interest in this type of volcanic activity was aroused by the 1912 Novarupta (Katmai) eruption, described by R. F. Griggs and by Fenner in a large number of papers spread over some thirty years (full bibliography in Fenner, 1950), and again by further eruptions of Mt. Pelée in the period 1929–1932. The latter were observed for a long period by Perret, who gave a full description of them and a full consideration of all the features of *nuées ardentes* (Perret, 1935). Interest has increased still further during the last few decades as a result of the increasing recognition, starting with P. Marshall's work in New Zealand, that this type of volcanic activity must have been very important in past geological periods, resulting in the extrusion of huge volumes of rocks of rhyolitic and andesitic composition.

Types

Although Anderson and Flett, as a result of their witness of a *nuée ardente*, questioned Lacroix's conclusion that flows of this type resulted from laterally directed explosions, it is now generally accepted that Lacroix was right concerning the flow that destroyed St. Pierre. Such lateral explosions are associated with the obstruction of the crater by a volcanic dome (tholoid) and/or the pipe by a spine. The extrusion of spines is a common late phase of volcanic activity that includes lateral explosions (for example, Mt. Pelée 1903). It is, however, accepted that *nuées ardentes* (including examples from Mt. Pelée similar to that witnessed by Anderson and Flett) can be of "overflowing" type emerging from an opening on the volcano without any accompanying explosion. Yet another type results from the gravitational collapse of part of an ascending column of pumice-rich material during a vulcanian- or plinian-type of eruption. This has sometimes been referred to as a base-surge, but this occurs at the commencement of an explosion and differs from the lateral surges resulting from gravitational collapse. The mechanics of the latter have been described by Sparks and Wilson (1976).

Nuées ardentes can also result from the collapse of a portion of an expanding volcanic dome, e.g., the 1930 eruption of Merapi (Java), and the Mount St. Helens (U.S.A.) 1980 eruption (Decker and Decker, 1981). The avalanching of hot, previously extruded ashes or lavas can also give rise to a similar phenomenon (Vesuvius in 1906), although generally on a smaller scale.

Features

If magma has the correct physical properties in relation to the structure of a volcano, then nuées ardentes may occur in rapid succession during a period of eruption, e.g., Perret (1935) recorded more than 150 from Mt. Pelée during the year 1930. These flows are capable of very rapid movement. Lacroix measured average velocities of two flows at 20 m/sec and 26.5 m/sec, while Perret measured one in 1929 at 33 m/sec and noted that for small flows velocities of 10–20 m/sec were common. He estimated that, in the destruction of St. Pierre in 1902, local velocities were as much as 150 m/sec. In that case, high overall velocities may have arisen from the combined effects of a laterally directed explosion and gravity, but locally very high velocities undoubtedly resulted from the rapid expansion of engulfed pockets of air. Other flows with lower mobility have traveled at much lower velocities and it has been estimated that the Valley of Ten Thousand Smokes deposit moved at only a few kilometers per hour.

The temperature of the nuée ardente that enveloped St. Pierre in 1902 was estimated to be about 800°C, while Perret concluded that a flow in 1930 had a temperature of about 1200°C at the moment of extrusion. From the characters of some pyroclastic flows it can be deduced that even higher temperatures are possible. The turbulent cloud of gases that rises from a nuée ardente may reach a height of several thousand feet.

The effect of gases in buoying up the fragments of liquid or solid material in a nuée ardente has long been recognized. Much of this gas may come from the continued escape of volatiles from the magma particles during the travel of the flow, but the rapid heating and expansion of air trapped in a flow can also help, and must be the main factor in the hot-ash avalanches mentioned earlier.

Origin

Ideas on the origin of nuées ardentes have often been involved in controversy and argument, even from the first recognition of this volcanic phenomenon in 1902. Lacroix advocated that the velocity of the nuée ardente that destroyed St. Pierre resulted from a lateral explosion, while Anderson and Flett claimed that the high velocity was caused by gravity acting on the almost frictionless flow. In that case, both views were correct, but it was quickly

realized that in some other eruptions gravity could be the only force leading to acceleration. The peléan type of eruption became recognized as a special category separate from a scale of increasing violence (Strombolian to Plinian). About the middle of this century, however, the idea began to arise that Peléan activity was the most violent type of explosive eruption. This resulted partly from a lack of precise definition of meanings and partly from the discovery of pyroclastic flow deposits associated with historic Plinian eruptions, e.g., Krakatoa 1883 and Katmai 1912. There was an enigma here, since very violent explosions ejecting material many kilometers into the atmosphere would separate the volcanic gases from the liquid and solid products of the eruption, yet peléan activity depends upon the suspension of the latter in gases. The solution came from the recognition that in some cases pyroclastic flows preceded the violently explosive phase (Katmai 1912, Mount St. Helens 1980), or quickly followed it (Krakatoa 1883), while in other eruptions nuées ardentes accompanied the Vulcanian type of activity through gravitational collapse of parts of the ascending column of material (Soufrière 1902, Mt. Lamington, Papua–New Guinea 1951).

W. E. TREMLETT

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Cross-references: *Explosive volcanic eruptions—classification*; *Pyroclastic eruptions*; *Pyroclastic flows*; *Volcanoes and volcanism*.

O

OCEAN-FLOOR METAMORPHISM

Ocean-floor metamorphism is a concept that has arisen from recent studies of present oceanic ridges and fracture zones where new crust is being generated, altered, and deformed. Recognition of on-land ophiolite suites as ancient examples of oceanic crust and mantle provide insights into the thermal and dynamothermal regimes that characterize ocean floors.

Metamorphism in modern oceanic crust

The most widely recognized variety of ocean-floor metamorphism is that resulting from burial and hydrothermal alteration in the high heat-flow and geothermal regime at mid-ocean ridges. New minerals are produced that show progressive changes with depth. The process is essentially a static hydration of original mineral assemblages without the production of a metamorphic mineral fabric in the affected rocks.

Metamorphic foliation in ocean-floor rocks is explained by directed stress at deeper levels of a spreading center or along oceanic fracture zones.

Burial metamorphism. Oceanic rocks metamorphosed in the zeolite, greenschist, and amphibolite facies have been dredged from oceanic ridge areas of considerable local relief. The metamorphic rocks typical of deeper crustal levels are exposed at the surface because of block faulting and vertical uplift. In a general way, the more metamorphosed rocks occur in deeper water, thus confirming vertical block faulting and an increase in metamorphic grade with depth. The consistency of rock type in the material dredged from a given depth below the tops of submarine scarps is indicative also of a systematically layered and metamorphosed oceanic crust prior to block faulting at the ridge.

Zeolite facies metamorphism of oceanic basalts takes place at very shallow depth, possibly beginning at a few tens of meters. Metamorphism in the greenschist facies begins at depths of about 1 km and it reaches the equivalent of amphibolite facies near the base of the crust at about 4–5 km depth. In zeolite facies metamorphism, plagioclase phenocrysts of oceanic rocks are altered to analcime and the

groundmass to other zeolites that include natrolite and saponite. In the greenschist facies the rocks still retain their igneous textures. Calcic plagioclase alters to albite, augite to actinolite, and olivine and glass to chlorite. In higher greenschist grades, epidote, tremolite, quartz, calcite, talc, and green hornblende also appear. Local dredge hauls of amphibolite facies metamorphic rocks indicate derivation from deep within layer 3 of the ocean floor. These rocks have lost their igneous textures and most relict minerals, and some exhibit a mineral lineation. Mineral assemblages include quartz, plagioclase (oligoclase–andesine), biotite, hornblende, epidote, diallagic diopside, sericitized orthoclase, magnetite, and sphene (Melson et al., 1968; Barrett and Aumento 1970; Aumento, 1972).

Figure 1 shows the consistent occurrence of metamorphosed basalts from depths >0.4 km and their absence at lesser depths. It also shows that less metamorphosed basalts occur in an uppermost layer of about 0.5 km in thickness. Shallow-dredge hauls yield mostly vesicular pillow basalts, some with zeolite replacement.

The junction between oceanic crustal layers 2 and 3 as defined by seismic wave velocities may reflect a metamorphic contrast. Layer 2 possibly consists of unmetamorphosed or zeolite facies mafic rocks, and layer 3 their more highly metamorphosed and denser greenschist–amphibolite facies equivalents.

Dynamothermal metamorphism. Deformed rocks of greenschist and amphibolite facies occur along transform faults and shear belts associated with spreading ridges. Greenschist facies metabasalts, basaltic mylonites, and amphibolites have been dredged from the major boundary faults of the Mid-Atlantic Rift Valley at 22°N latitude, the Vema Fracture zone and Gorringer Bank. Greenschists and amphibolites that constitute the basement to some island arcs, and show rapid changes in the intensity of deformation, may have formed in a similar way, or by tectonic processes that gave rise to the arc volcanic piles (e.g., Melson et al., 1968; Honnorez et al., 1984).

Subduction, or the descent of oceanic crust along downward-dipping zones, normally leads to the formation of high-pressure facies mineral assemblages that include glaucophane, lawson-

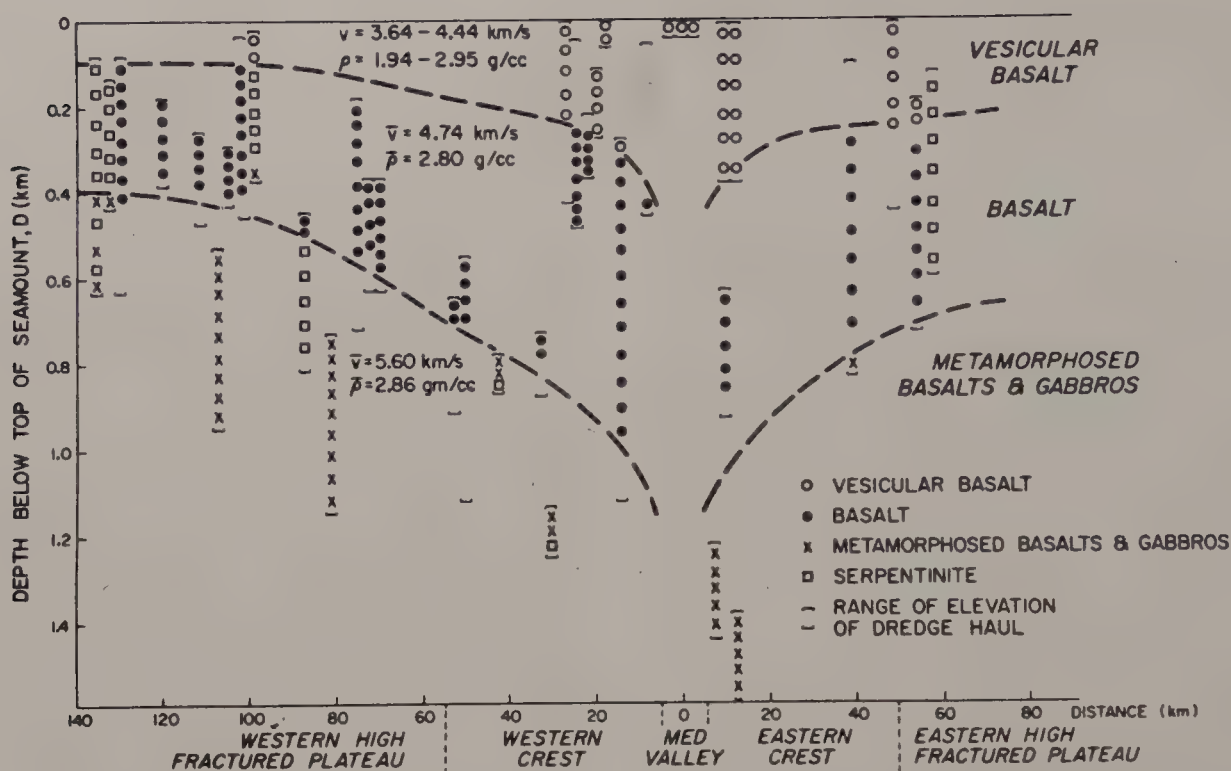


FIGURE 1. Depth of dredge hauls below the tops of submarine scarps plotted against distance from the Mid-Atlantic Ridge axis. (From Barrett and Aumento, 1970.)

ite, jadeite, and pyrope garnet. These usually occur in the outer or oceanward metamorphic belts of modern island arcs, e.g., Japan.

At mantle depths, ductile deformation that accompanies recrystallization may relate to convective mass transfer of dense semiconsolidated material. Mantle-derived material may be further deformed or mylonitized during its solid-state ascent to the surface. Ultramafic rocks dredged from the Mid-Atlantic Ridge at 45°N latitude show evidence of mylonitization and other mechanical deformation (Aumento, 1972). Similarly, at St. Paul's Rocks on the equatorial Mid-Atlantic Ridge, peridotite, brown hornblende-rich ultrabasic alkaline rocks, and gabbro are mylonitized. This occurrence of deformed ultramafic rocks is unique, for most ridge islands are volcanic. Mineral assemblages in the ultramafic rocks indicate equilibrium parageneses compatible with upper mantle depths. Mylonitization is attributed to later solid-state emplacement at much lower temperatures and pressures (Melson et al., 1972).

Metamorphism in ancient oceanic rocks

The interpretation of the ophiolite suite of rock units as on-land oceanic crust and mantle affords an opportunity to compare the effects of ocean-floor metamorphism in ancient and modern oceanic crust. It also provides an

opportunity to study levels of the crust and mantle in the ancient examples to depths beyond the penetration of dredging or drilling in modern oceans.

Burial metamorphism. The Cretaceous Troodos ophiolite complex of Cyprus has zeolite facies mineral assemblages in its basaltic upper layers and greenschist metamorphism at the level of sheeted dykes. A progressive depth-controlled distribution of zeolite minerals (Fig. 2) may be a simplification of more complex burial and hydrothermal effects.

A break in the increasing temperature of zeolite facies metamorphism may reflect an unconformity between upper and lower pillow lavas. The zeolite-greenschist facies boundary may be either transitional or sharp and an intervening prehnite-pumpellyite facies zone found both in other ophiolite complexes (e.g., Bay of Islands Complex—Williams and Malpas, 1972), and some oceanic rocks (Melson et al., 1968), is absent. Indirect geological reasoning suggests that ocean-floor burial metamorphism was imprinted upon the Troodos rocks within 100 km of the oceanic ridge spreading axis, and that it took place at a constant distance from the ridge throughout the spreading process. Deeper gabbroic levels of the Troodos ophiolite suite are unmetamorphosed.

Troodos greenschists have a rigidity and seismic velocity that are distinctly greater than

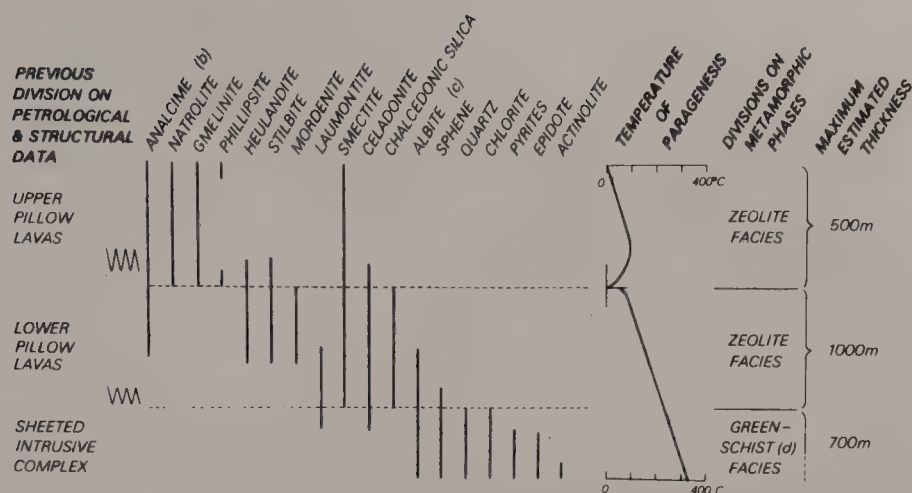


FIGURE 2. Metamorphic minerals and facies and their depth-controlled distribution within the Troodos massif, Cyprus. (From Gass and Smewing, 1973.)

those of rocks in the overlying zeolite facies zone. The zeolite–greenschist facies boundary in the ancient example could therefore mark the seismic discontinuity between oceanic crustal layers 2 and 3, as has also been suggested for modern oceanic crust at the Mid-Atlantic Ridge.

The Early Palaeozoic Bay of Islands ophiolite complex in Newfoundland generally has prehnite–pumpellyite facies metamorphism in its basaltic rocks. Pumpellyite decreases in abundance with depth, so that at the sheeted dyke level and upper parts of the underlying gabbro level, greenschist to low-amphibolite facies mineral assemblages are predominant. Local areas with zeolite minerals (mainly laumontite) at the top of the ophiolite sequence are interpreted as erosional remnants of a zeolite facies zone that once existed above the prehnite–pumpellyite zone. Deeper levels of the gabbro unit are relatively unmetamorphosed, possibly because of a lack of surficial hydrous fluids necessary to accomplish static hydration. The water necessary for hydration in burial metamorphism of this kind is inferred to be sea water that permeates higher parts of the crustal section along cracks in distending ridges and moves through the rocks by diffusion over short distances (Williams and Malpas, 1972). Oxygen isotope studies on oceanic rocks support this suggestion (Spooner et al., 1974; Gregory and Taylor, 1981).

Thermal (contact) metamorphism. Small quartz diorite and gabbro intrusions in some ancient ophiolite complexes have thin aureoles characterized by a gradual outward transition from blue-green hornblende through actinolite to chlorite, all found within 30 m of the intrusive contact and with preexisting textures preserved (Malpas et al., 1973). Quartz diorites, trondhjem-

ites, and albite granites similar to the small intrusions in ancient ophiolite complexes have been dredged from modern ridges (Aumento, 1972), so that thermal (contact) metamorphism may also be a local phenomenon there (Fig. 3). The metamorphic features exhibited throughout a complete ophiolite section, as illustrated by the Bay of Islands Complex, Newfoundland.

Dynamothermal metamorphism. Foliated metamorphic rocks are restricted to deeper levels of ophiolite complexes. Conformable amphibolite zones in gabbro indicate ductile deformation, possibly related to horizontal transport away from the spreading center. Recrystallized foliated ultramafic rocks form a thick zone still deeper beneath the cumulate rocks. The foliated ultramafic rocks, which are equivalent to the mantle beneath present oceanic crust, are folded and they are cut by massive pyroxenite dykes. Because the pyroxenite dykes are indigenous to the ophiolite suite, the earlier metamorphism and deformation must have taken place deep in the mantle, i.e., the ultramafic rocks are mantle tectonites.

Greenschists, amphibolites, and foliated sodic granites of the ophiolitic Little Port Complex (Williams, 1973) in western Newfoundland are interpreted as Bay of Islands equivalents that were deformed and metamorphosed along a Palaeozoic oceanic fracture zone (Karson, 1984).

Thin dynamothermal metamorphic soles at the stratigraphic base of ophiolite suites and associated with isolated masses of alpine-type peridotite are interpreted as the result of forceful expulsion and transport of hot mantle and oceanic crust across supracrustal rocks (Williams and Smyth, 1973; Treloar et al., 1980). These sole rocks show decreasing metamorphic grade and decreasing intensity of

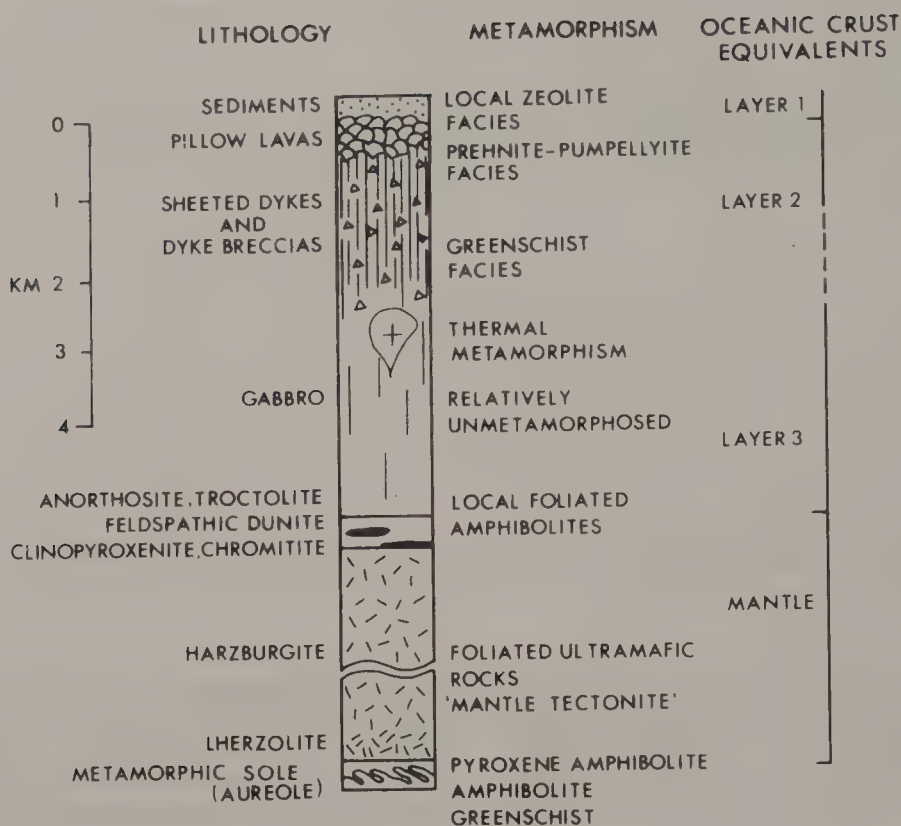


FIGURE 3. Metamorphic features of the Bay of Islands Complex.

deformation downward and away from the base of the ophiolite suite. Immediately overlying ultramafic rocks that contain deep brown amphibole, clinopyroxene, garnet, and phlogopite are interpreted as peridotites metamorphosed during the same transport process, rather than primary mantle assemblages.

HAROLD WILLIAMS

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Cross-references: *Amphibolite facies*; *Aureole*; *Brownstone facies*; *Burial metamorphism*; *Greenschist facies*; *Mid-ocean ridge and ocean-floor petrology*; *Ophiolite*; *Zeolite facies*.

OCEAN-ISLAND IGNEOUS ROCKS

Ocean islands are volcanic edifices built up from the ocean floor, and such edifices include active volcanic seamounts that have not yet reached the ocean surface, major active volcanic islands, and extinct volcanic islands now partially or totally submerged by isostatic adjustments of the ocean floor. In this broad grouping are included islands like Hawaii and Reunion, active seamounts like Macdonald, and extinct islands and seamounts (guyots) such as those in the northwestern extension of the Hawaiian island chain.

The presence of such volcanic centers away from plate margins is considered to reflect the existence of mantle plumes or hot spots, and distinctive suites of igneous rocks are the result. Variations arise from the location of some hot spots coincident with mid-ocean ridges—e.g., Iceland—or where hot spot migration has crossed or is located on the continent–ocean transition zone—e.g., the Canary Islands and the islands in the Gulf of Guinea.

The bulk of volcanic material in an ocean island is tholeiitic basalt, but more differentiated rocks are also present (Fig. 1). A number of distinct suites are recognized: (1) tholeiitic basalt—silica-oversaturated peralkaline rocks, (2) alkaline basalts—trachyte and phonolite (mildly undersaturated), and (3) olivine basalt—nephelinites (strongly undersaturated).

Tholeiitic basalts to silica-oversaturated peralkaline rocks represent a bimodal suite in which both saturated and oversaturated basalts are present, i.e., hypersthene- and hypersthene-quartz-normative rocks. The second modal group consists of alkali-rich (Na and K) quartz-normative felsic rocks, variously known as *alkali rhyolite*, *comendite*, and *pantellerite*, and the group known as *icelandites*, that have also been called *oceanic andesites*. Although superficially similar to members of the calc-alkaline suite, they are chemically and mineralogically distinct (total alkali:Al ratios) and the icelandites are probably hybrids.

The *alkaline basalt* to *trachyte* and *phonolite* suite is continuous through trachybasalt and intermediate members such as *mugearite* and

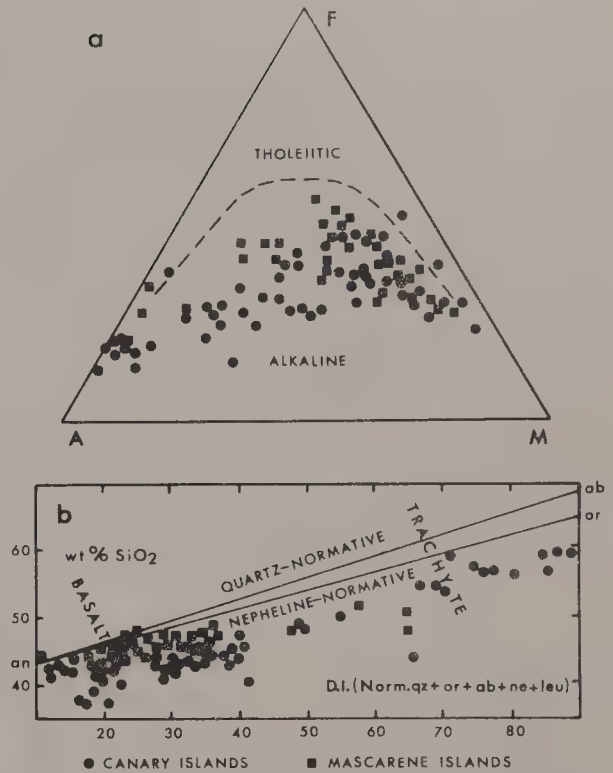


FIGURE 1. Chemical variation in alkaline volcanic rocks from ocean islands—Canary Islands and Mascarene Islands (Mauritius, Réunion, and Rodriguez). (a) AFM plot showing segregation of tholeiitic and alkaline varieties. (b) SiO_2 vs. differentiation index (DI) showing segregation between quartz-normative and nepheline-normative rocks in the basalt–trachyte suite. (After Borley, 1973.)

hawaiite. They are all usually nepheline- and/or leucite-normative, but rarely carry modal feldspathoids when they are eruptive. Their intrusive equivalents may contain nepheline.

Olivine basalts and *nephelinites* are members of a suite that includes strongly undersaturated rocks, which may also show a very broad variation in color index, including both felsic and mafic or ultramafic members. Undersaturation is such that both normative and modal nepheline are present, and minerals such as melilite and sodalite may occur. Members include basanite and tephrite, and the mafic olivine basaltic rocks ankaramite and ankaramite, through the oceanites to rocks like limburgite and nephelinite itself. The intrusive equivalents are essexite and ijolite.

While the bulk of most ocean islands is tholeiitic basalt, the differentiated rocks often include members of all three suites, but the proportions vary widely. The rocks of the bimodal tholeiitic basalt suite are particularly abundant on those islands that straddle or are close to, but offset from, mid-ocean spreading ridges—e.g., Iceland, Azores, Easter Island.

The most usual plan of an ocean island involves a shield volcano consisting mainly of

basalt, capped by domes and cones of the more chemically evolved rocks. Thus, the shield volcanoes of the Hawaiian islands consist of >99% basalt, with the undersaturated trachytes, phonolites and nephelinites restricted to parasitic cones and lava domes, whereas Gran Canaria's shield is crowned by the 2000 m high basalt-phonolite-trachyte cone of the Teide volcano. In contrast, almost the entire edifice of Trinidad appears to consist of undersaturated and feldspathoidal rocks.

On the Hawaiian islands there is a secular separation of the basaltic to trachytic-phonolitic activity and the eruption of the nephelinite suite, marked by a period of erosion. In the Canary Islands a less marked hiatus separates the early basaltic shield formation from the more undersaturated stratovolcanoes.

Eruption style

Oceanic basalt shield volcanoes carry some form of summit crater, or major caldera. Eruptions may be both localized at the summit, or may issue from dilational fissures in well-defined rift zones that may radiate from the central craters. On Hawaii activity can occur at both sites simultaneously, with lava lakes accumulating in the calderas of Mauna Loa and Kilauea, drained by eruptions from the rift zones (Fig. 2a). The sites of fissure eruptions are often marked by lines of cinder cones. On Hawaii the onset of an eruption cycle is marked by the inflation of the summit area followed by dilation of the rift zone before an effusive outbreak. Similar patterns of activity are recorded from Iceland, where the major cauldrons represent centers for magma ascent within dilational fissure swarms marking the onshore extension of the Mid-Atlantic spreading ridge.

As the volcanic edifice grows from the ocean floor, the style of eruption and effusive products undergo changes which relate to the changing environment of eruption (water depth, particularly) and the evolving magma composition. When eruptions occur at water depths where hydrostatic pressure suppresses steam generation, lava flows and pillows are the only effusive products (Fig. 2b). Once the edifice rises above a critical depth, which depends on the gas pressure of the lava and its temperature, a more explosive, phreatic activity generates hyalotuff and autoclastic breccia that is interbedded with more massive flows and pillows. This interlayering marks the whole shallow-water zone, and is capped by the subaerial edifice, which consists of massive lava flows with thin, interbedded pyroclastic units—air-fall and fluvially reworked tuffs. Where subaerial flows

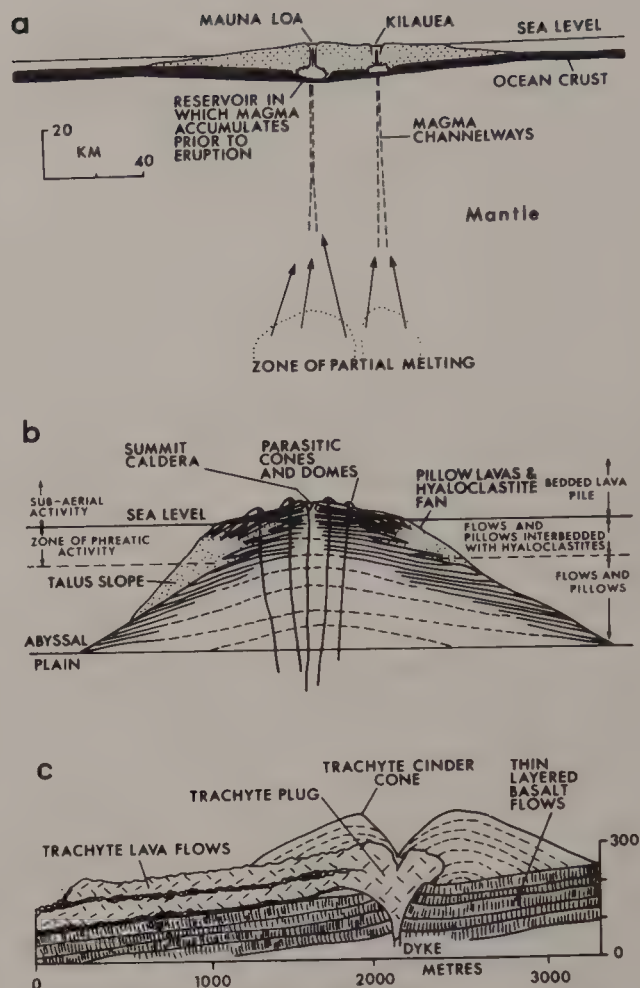


FIGURE 2. (a) True scale section of the island of Hawaii showing the shield form of the volcano, with the depression of the ocean crust at the crest of the Hawaiian rise. (After Macdonald and Abbott, 1970.) (b) Schematic (not to scale) section through an ocean island showing changes of effusive products as the edifice grows into shallow water and the magma type changes. (After Williams and McBirney, 1979.) (c) Section through trachyte cinder cone, trachyte plug and trachyte lava flows, Hawaii. (After Macdonald and Abbot, 1970.)

reach the sea, hyaloclastite deltas are built out, capped by cogenetic flows and mixed with cogenetic pillow lavas fed by lava tubes. Likewise, as lava composition changes, the quiet but voluminous basalt effusion gives way to less voluminous, but more explosive trachytic, phonolitic, or nephelinitic eruptions (Fig. 2c).

Hot spots

“Hot spots” on the Earth’s surface, marked by within-plate volcanic centers are considered to overlie an upwelling plume in the deep mantle. The chemistry of ocean-island basalts,

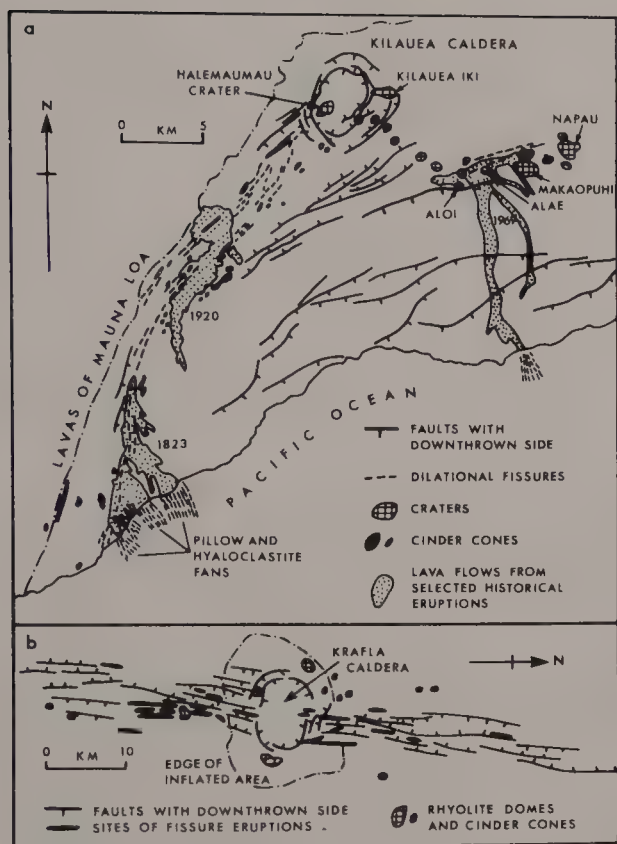


FIGURE 3. Calderas and rifts on ocean islands. (a) Kilauea caldera and the western rift zone, Hawaii, showing selected recent lava flows. (After Macdonald and Abbott, 1970.) (b) Krafla caldera and the dilational fissure swarm Iceland, showing sites of fissure eruptions. (After Saemundsson, 1978.)

consistent with partial melting of mantle peridotite enriched in many trace elements when compared to the source region for mid-ocean ridge basalt (MORB), fits such a model. Likewise, the position of these plumes is reflected by gravity anomalies associated with elevated regions of the ocean floor, e.g., the Hawaiian ridge and the Iceland–Faroe rise (Fig. 3). In the Pacific Ocean the relative motion of the oceanic lithosphere over a hot spot is considered to explain the presence of linear arrays of volcanic islands, e.g., the Hawaiian island chain, Line Island chain, and the Tuamotu archipelago, the Marshall–Gilbert and Ellice Island chain, and the Austral Island chain. In each case the active volcanism is sited at the southeastern end of the line, and in the Hawaiian chain, and its continuation in the Emperor seamounts, the volcanic edifices become older to the northwest (Fig. 4). Each of the three alignments mentioned here includes a bend that may reflect a change in the movement direction of the Pacific plate.

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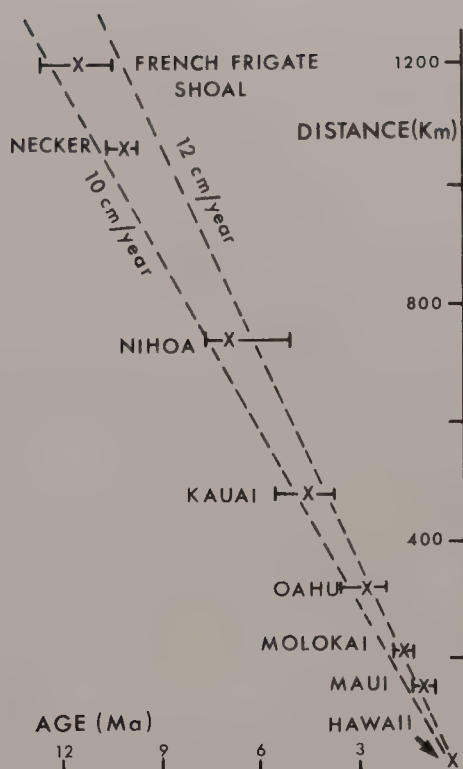


FIGURE 4. K–Ar age dates for volcanic centers in the Hawaiian chain showing how the centers get older away from Hawaii (the currently active site) consistent with a rate of migration related to motion of the Pacific plates of 10–12 cm/yr. (After Dalrymple et al., 1973, in Williams and McBirney, 1979.)

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Cross-references: *Alkaline rocks—undersaturated; Basalt; Igneous rocks—tectonic setting; Mid-ocean ridge and ocean-floor petrology; Nephelinite; Peralkaline igneous rocks; Phonolite; Rhyolite; Trachyte; Volcanic rocks; Volcanoes and volcanism.*

OCEANIC CRUST

The term “crust” originally referred to what was considered to be a hard crystalline shell of

the Earth covering a generally molten interior. With the development of instrumental seismology, the Earth's crust took on a new definition. It was found that a major seismic boundary, known as the Mohorovičić discontinuity (Moho), is present at depths of ca. 10 km below sea level in the oceans and at depths of 30–50 km below the continents. Above this boundary, measured velocities of compressional waves are usually below 7.5 km/sec, whereas immediately below the Moho compressional wave velocities are about 8 km/sec. The Earth's crust is separated from the underlying mantle by this discontinuity. The sharpness of the boundary is uncertain; however, it is widely considered that the transition from the lower crust to the mantle takes place over a fairly narrow depth interval of a few kilometers or less.

Sea-floor spreading and the oceanic crust

During the past three decades, the interpretation of geomagnetic data and sediment ages from deep-sea drilling have provided powerful support for the hypothesis of continental drift and sea-floor spreading. New oceanic crust is continuously being formed along the crest of a worldwide system of ridges (Fig. 1). Once formed, the oceanic crust is considered to be

transported laterally away from the ridge systems as part of a system of rigid plates (Le Pichon, 1968) which extend vertically from the surface to depths of 70–100 km. The leading edges of the plates are destroyed at oceanic trenches by downward movement into the mantle. These rigid plates, which include the crust and uppermost mantle, form what is termed the *lithosphere*. Below the lithosphere to a depth of several hundred kilometers is a region with much smaller yield strength. The upper portion of this region, the *asthenosphere*, corresponds to the low-velocity layer of seismology.

Structure of the oceanic crust

The principal sources of information about the oceanic crust come from geophysical data. Heat flow, gravity, magnetism, and seismology have all provided important data on the nature of the oceanic crust. During the past three decades, information accumulated from seismic refraction techniques on oceanic crustal structure has shown that the oceanic crust possesses a layered structure that is relatively simple compared with continental crustal structure. Seismic refraction measurements usually divide the oceanic crust into three layers. These three layers are observed in almost all major oceanic



FIGURE 1. Locations of mid-oceanic ridges (thick lines), fracture zones (thin lines which offset the ridge system) and the six major plates of the lithosphere. (After Le Pichon, 1968.)

TABLE 1. Average oceanic crustal structure

Layer	Thickness (km)	Velocity (km/sec)
1	0.45	2.0
2	1.71 ± 0.75	5.07 ± 0.63
3	4.86 ± 1.42	6.69 ± 0.26

Source: after Raitt (1963).

regions. In addition to layering, seismic refraction measurements provide information on the elastic properties (usually in the form of compressional wave velocities from first arrivals of seismic energy) and thicknesses of the individual layers. Averages of these properties and their standard deviations for over 100 deep-sea seismic refraction stations from the Atlantic, Pacific, and Indian oceans are given in Table 1.

Layer 1. The uppermost layer of the oceanic crust consists of unconsolidated or semiconsolidated sediments. Its thickness varies from zero on topographic highs to several kilometers in some oceanic deeps and continental rise areas. Average thickness appears to be a few hundred meters. Sediments of layer 1 bordering the continents are generally classified as terrigenous, meaning that they are dominantly made up of material transported from the continents. Deep-sea sediments of layer 1 consist primarily of calcareous and siliceous remains of marine organisms and red clay. Drilling has demonstrated that many clay minerals, carbonates, and zeolites form chemically in layer 1 after deposition of these deep-sea sediments. In addition, deep-sea sediments are sometimes composed of volcanic ash and terrigenous sediments that have been transported from the continental margins by slumping and turbidity currents.

Detailed seismic studies of layer 1 are complicated by poor acoustic properties of the sediments (Ewing and Nafe, 1963). In many localities the velocities of compressional waves in the upper part of the sediments are less than or equal to seawater velocity. In the upper few hundred meters of layer 1 there are appreciable velocity gradients that average between 0.9 and 1.4 km/sec. The high gradients are probably related to rapid decreases in porosity. At greater depths lower gradients are observed that suggest a decrease in the rate of compaction of the sediments (Ewing and Nafe, 1963).

Results from the Deep Sea Drilling Project (DSDP) and the Ocean Drilling Project (ODP) have established that the palaeontological ages of sediments at the base of layer 1 correspond very closely with age predictions based on magnetic anomalies. Thus, the ages of basal

sediments in layer 1 increase away from centers of sea-floor spreading. Ages of oldest sediments recovered immediately above layer 2 range from a few million years at ridge crests to 140 Ma in the NW Pacific Ocean.

Layer 2. Layer 2 has frequently been referred to as "basement." It is characterized by higher compressional wave velocities (4.0–6.0 km/sec) than the overlying sediments of layer 1. Seismic studies are complicated by the small distance over which refracted first arrivals are received from layer 2 and the irregular interface which often exists between layers 1 and 2 (Raitt, 1963). These difficulties could be partially responsible for the wide range in velocities reported for layer 2. Strong velocity gradients have been observed in layer 2 (Houtz and Ewing, 1976), which have been interpreted to originate from decreasing porosity with depth. It appears probable that much of the variability in velocity reported for layer 2 is related to porosity. Oriented cracks may be responsible for observed seismic anisotropy in layer 2 (Stephen, 1985).

Drilling in the Atlantic and Pacific ocean basins through the overlying sediment cover and into the top few meters of layer 2 has shown that the uppermost portion of layer 2 consists of basalt and altered basalt. The presence of basalt in layer 2 is also supported by dredgings over the oceanic ridge systems where the sediment cover is thin or absent. Although alkali basalts have been sampled near ridge crests, the dominant basalt appears to be tholeiitic. The exact nature of the lower portion of layer 2 is presently unknown. However, based on ophiolite models for the oceanic crust and drilling near the Costa Rica Rift, it is probable that in many regions the lower portion of layer 2 consists of sheeted dykes of dolerite.

Layer 3. Layer 3 or the "oceanic layer" constitutes over two-thirds by volume of the oceanic crust. Seismic refraction observations of layer 3 are generally better than those of layer 2 because first arrivals from layer 3 are received over a relatively large range of shot distances. Average velocities and thicknesses of layer 3 for different oceanic areas are given in Table 2 (data tabulated by Raitt (1963), with the exception of more recent data included in the Indian Ocean averages). Several studies have found evidence for fine structure within layer 3 (Christensen and Salisbury, 1975; Spudich and Orcutt, 1980). In some regions, velocities apparently increase with depth and in others they decrease.

Crustal structure over ocean ridges

The concepts of continental drift, sea-floor spreading, and plate tectonics lead to considera-

TABLE 2. Regional seismic structure of layer 3

	Velocity (km/sec)			Thickness (km)		
	No.	Mean	S.D.	No.	Mean	S.D.
Western N Atlantic Ocean	65	6.70	0.27	38	4.76	1.41
Eastern N Atlantic Ocean	13	6.56	0.32	7	4.41	1.03
Eastern N Pacific Ocean	11	6.75	0.17	7	4.64	0.79
Western N Pacific Ocean	16	6.73	0.25	9	5.08	1.72
S Pacific Ocean	12	6.72	0.20	11	5.43	1.72
Indian Ocean	16	6.74	0.25	8	3.84	0.88

tion of the oceanic crust, to the problems of its constitution and of the chemical and physical processes that account for its generation. It is generally accepted that oceanic crust is generated in the vicinity of oceanic ridges. The oceanic ridges form a world-encircling mountain range cut into sections by fracture zones (Fig. 1). This mountain range on the sea floor appears to differ in various ways from continental mountain ranges. First, exploration of the sea floor over the oceanic ridges shows that on the surface they are composed primarily of basic volcanic rocks. Dredgings along large fault scarps that form a prominent median valley for most of the length of the Mid-Atlantic Ridge system suggest that the surficial basalt may be metamorphosed to low-grade greenschist facies rocks at shallow depths under the ridges. The mountain ranges of continental regions, on the other hand, are primarily composed of sedimentary rocks and their metamorphosed equivalents. Second, the rock of the ocean-ridge system show little deformation by compressive stress, whereas rocks from the mountains of continents are usually highly folded.

Seismic refraction work and gravity measure-

ments have demonstrated that oceanic crust under mid-oceanic ridge crests differs significantly from oceanic crust under the ocean basins. Oceanic crustal models across the northern Mid-Atlantic Ridge and E Pacific Rise determined by Talwani et al. (1965) from gravity and seismic refraction data are shown in Fig. 2. Sediment cover (layer 1) over the ridges tends to be thin or even absent in many regions. Under both ridges there are thick sections of layer 2 (velocities of 4.5–5.8 km/sec) that extend across the crests. The major difference in crustal structure under ridge crests and ocean basins is the systematic decrease in total crustal thickness under the ridges.

Below layer 3 of the E Pacific Rise and layer 2 of the Mid-Atlantic Ridge, compressional wave velocities range from 7.3 to 7.6 km/sec. These velocities are generally higher than oceanic crustal velocities and lower than velocities (8.1 km/sec) observed for the upper mantle along the flanks of the ocean ridges and beneath normal oceanic crust. This region below ridge crests has been referred to as the 7.3 km/sec layer, altered mantle, and anomalous mantle. Crustal models calculated from gravity and seismic data suggest that this region extends beneath the normal mantle under the flanks of the mid-ocean ridges (Talwani et al., 1965). The origin of this anomalous region is not completely clear; however, it appears probable that high temperatures associated with high heat flow and accompanying partial melting could be, at least in part, responsible for the low mantle velocities. This region most likely plays an important role in the formation of oceanic crust.

Rocks of the oceanic crust

Several different rock types have been obtained from the ocean floor by dredging and the recovered rocks have frequently been cited as evidence supporting various petrologic models of the oceanic crust. Ultramafic and mafic rocks have been by far the most abundant igneous and metamorphic rocks recovered.

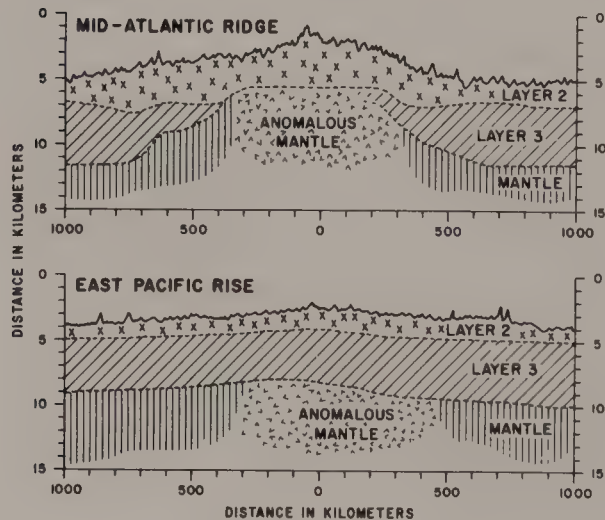


FIGURE 2. Structure across the northern Mid-Atlantic Ridge and E Pacific Rise; sediment cover (layer 1) not shown. (After Talwani et al., 1965.)

Favored locations for dredging include ridge crests where sediment cover is thin or absent and major fracture zones where the relief is often several kilometers. A greater variety of rock has been reported from fracture zones because deeper portions of the crust are exposed along their walls.

Basalts are generally dredged from ridge crests and the upper slopes of fracture zones. Although some systematic compositional variations are apparent, the chemistry and mineralogy of most oceanic basalts are consistent with low-K olivine tholeiites. Basalts with alkaline affinities have been reported but appear quite rare. Frequently sampled with the basalts are coarser-grained (0.5–2.0 mm) equivalents referred to as dolerites. Other mafic rocks usually recovered from deeper dredge hauls within fracture zones include gabbros, norites, amphibolites, greenstones, and greenschists. The metabasalts vary from well-foliated to granular.

Ultramafic rocks from the oceanic crust consist of peridotite, partially serpentinized peridotite, and serpentinite. Olivine and enstatite are the most abundant minerals in the peridotites, and additional minerals frequently reported are clinopyroxene, plagioclase, and Cr-spinel. Many of the peridotites have typical textures of metamorphic tectonites, showing abundant shear banding and granulation. Serpentinization is highly variable, but is often related to mechanical deformation. Lizardite and chrysotile are the most abundant serpentine minerals.

Composition of layers 2 and 3

The classical model for the composition of the continental crust, based largely on early petrological investigations of magmatic differentiation, consisted of a granitic upper portion overlying basalt or gabbro. Since seismic velocities in layer 3 were found to be similar to many portions of the lower continental crust, it was generally assumed that the lower oceanic crust was composed of gabbro or basalt. In 1962, Hess proposed that layer 3

consisted of peridotite 70% serpentinized. Another possibility (for references see Christensen and Salisbury, 1975) is that the lower oceanic crust is composed of gabbro and metamorphosed basalt and gabbro. The implications of these oceanic crustal models differ sufficiently significantly that progress in understanding the processes of crustal formation in the vicinity of ridge crests requires a definite choice between them.

Recent laboratory measurements of the velocities of rocks obtained from oceanic regions (Table 3) provide important information as to possible mineral assemblages of the oceanic crustal layers and, perhaps more important, may aid in rejecting many rocks as major crustal constituents. Comparisons of this type must necessarily assume a certain degree of homogeneity in the crustal layers, at least over distances comparable to the path lengths of waves used in seismic refraction investigations.

Laboratory measurements of velocities of compressional waves in samples of basalt, altered basalt, and dolerite dredged from the Mid-Atlantic Ridge are in the range of velocities observed for layer 2. This is consistent with drilling the ocean basins, which has recovered basic volcanic rocks from the upper portion of layer 2, and models of the oceanic crust based on ophiolites. The wide range of velocities reported for layer 2 are probably related to porosity and alteration, perhaps manifested by initial stages of metamorphism that increases with depth in layer 2.

Velocities at appropriate pressures (1–2 kb) measured for oceanic volcanic rocks and serpentinites (Christensen, 1970, 1972) are significantly lower than layer 3 velocities (Table 3). Thus, it appears that basalt and serpentinite are not major constituents of layer 3. Compressional wave velocities in gabbro, partially serpentinized peridotite, relatively high-grade greenstone (Fig. 3), and amphibolite are in the range of velocities reported for layer 3. Even though these limitations can be placed on the composition of layer 3, its exact petrologic nature will remain open to speculation until

TABLE 3. Average compressional wave velocities (km/sec) as a function of pressure for rocks dredged from oceanic ridges

Rock	No.	Location	P (kb)			
			0.5	1.0	2.0	4.0
Tholeiitic basalt	11	Mid-Atlantic Ridge	5.79	5.96	6.14	6.31
Altered basalt	17	Mid-Atlantic ridge	5.71	5.88	6.08	6.29
Tholeiitic basalt	8	Juan de Fuca Ridge	4.70	5.41	5.97	6.27
Dolerite	9	Mid-Atlantic Ridge	6.03	6.09	6.20	6.34
Serpentinite	9	Mid-Atlantic Ridge	4.69	4.82	4.98	5.18

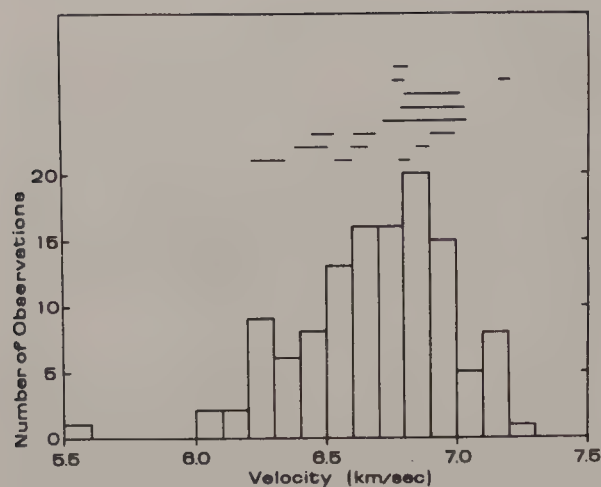


FIGURE 3. Histogram of compressional wave velocities reported for layer 3; horizontal bars represent velocities measured at pressures of 1–2 kb for greenstones.

direct sampling is accomplished by deep crustal drilling (Christensen and Salisbury, 1975).

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Cross-references: *Basalt*; *Mid-ocean ridge and ocean-floor petrology*; *Ocean-floor metamorphism*; *Ophiolite*; *Peridotite*; *Serpentine*.

OPHIOLITE

The term *ophiolite* derives from the Greek *ophis* (snake) referring to the similarity in appearance between altered ultramafic rocks (serpentinites) and snake skin (from Latin *serpentium*, via Medieval Latin *serpentin*). In the modern sense (for reviews of historical concepts see **Alpine mafic magma stem**; **Steinmann Trinity**) an ophiolite is an assemblage of rock types whose petrogenetic and paragenetic characteristics indicate an origin through processes comparable to those operative at mid-ocean ridges. Ophiolites are allochthonous units now located in fold mountain belts (Fig. 1) and are regarded as translated fragments of oceanic crust and underlying lithosphere and mantle.

The ideal ophiolite should consist of the following assemblages, which are related to oceanic seismic layers.

- **Layer 1. Sedimentary unit** consisting of pelagic deposits, e.g., radiolarian chert, calcilutite, red argillites, minor volcanic material.
- **Layer 2. Lava unit** consisting predominantly of pillow basalt that is often spilitized, with interstitial palagonite and/or pelagic sediment; dykes, sills, lava tubes, and massive flows may be present.
Sheeted dyke complex grading upwards into the above unit, where screens of lava separate dykes; in the core of this layer the unit consists of 100% dyke—usually quartz dolerite with minor gabbro, albitite, and rodingite.
- **Layer 3. Isotropic gabbro unit** grading down from where gabbro screens separate dolerite dykes, and gabbro dykes interleave with dolerite, to the major part of this unit that consists of massive quartz gabbro, gabbro or troctolite with minor diorite, quartz diorite, trondhjemite, plagiogranite, or granophyre.
Layered cumulate unit consisting of cumulate gabbros, and cumulate layers of dunite, pyroxenite and anorthosite.
- **Layer 4—mantle. Ultramafic tectonite unit** consisting of anisotropic harzburgite, often with mylonitic fabric, and variably serpentinitized.
Massive ultramafic unit consisting of variously serpentinitized harzburgite, lherzolite, dunite and pyroxenites.
Tectonic break—basal tectonite often underlying the whole assemblage; this contains low-*T*–high-*P* (glaucophane schist facies) mineral assemblages or high-*T*–high-*P* (eclogite facies) mineral assemblages (thermal aureole). It may include a tectonic mélange containing megacrysts from the other ophiolite



FIGURE 1. Spatial and temporal distribution of ophiolites in fold mountain belts.

lite units and represents the "sole" on which the ophiolite complex was transported.

This ideal ophiolite is rarely preserved in any one complex: one or more of the units is usually missing. Ophiolites are usually contained either in distinct thrust slices (nappes), or as enclaves within a larger allochthonous unit, or as blocks in mélangé or olistostrome. More rarely they are still attached to in situ back-arc basin crust or ocean floor.

Examples

Bay of Islands Complex, Newfoundland. This is an Ordovician ophiolite in the Appalachian fold belt of W Newfoundland (Williams, 1973). All the units of an ideal ophiolite have been recognized, while the nature of the basal tectonite and its inverted thermal aureole were first recognized here (Williams and Smyth, 1973). Related ophiolites in Newfoundland include those of the Baie Verte zone, considered to represent a small back-arc basin (Kidd, 1977), and the Coastal Complex, now regarded as an oceanic fracture zone assemblage (Karson and Dewey, 1978). The ophiolite complex itself is preserved as nappe klippen that overrode an Ordovician sedimentary succession including olistostromes.

Semail ophiolite, Oman. This late Cretaceous ophiolite forms the eastern end of a belt of such complexes extending through the Zagros fold belt of Iran, Iraq and E Turkey to Cyprus. All the units of an ideal ophiolite are present somewhere in the 350 km-long Oman ophiolite, though complete sequences in any one sector

are not usually preserved. The complex forms an "ophiolite nappe" emplaced onto a Jurassic-Cretaceous continental shelf, and underlain by both basal tectonite, with an inverted aureole, and an olistostrome that contains ophiolite blocks. The volcanic rocks of layer 2a and layer 1 include the products of island arc-type volcanism, superimposed on an ocean-floor succession (Pearce et al., 1981).

Troodos Complex, Cyprus. This is one of the best-documented ophiolites (Fig. 2), lying at the western end of the belt that includes the Semail nappe of Oman. All the layers of a model ophiolite are present, though the base is not exposed. The ophiolite itself is little modified by post-obduction processes and tectonism, so that the relationship between the various layers is quite clear. An extensive

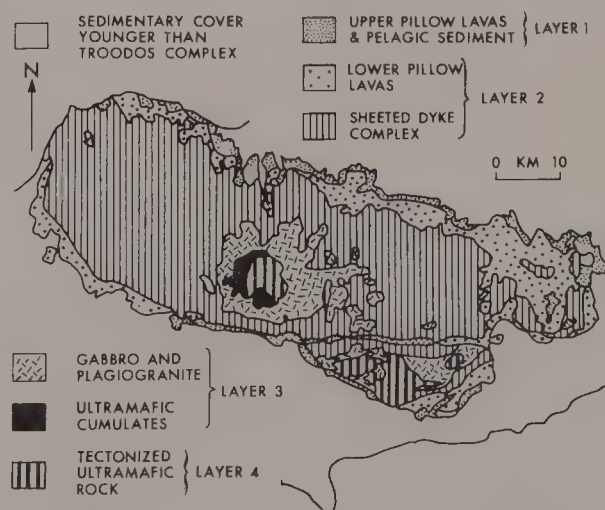


FIGURE 2. Map of the Troodos Complex, Cyprus.

sheeted dyke complex is very well preserved, and its study proved seminal in the understanding of mid-ocean spreading ridge processes (see **Mid-ocean ridge and ocean-floor petrology**). The volcanic rocks and sediments of ocean layers 1 and 2 are exceptionally well preserved, and include examples of off-axis high-Mg basalts and picrites, as well as radiolarian cherts and metalliferous muds (umbers). The economic importance of this ophiolite has been appreciated since antiquity, because of the abundance of Cu ores (Cyprus derives its name from *cuprios*, Greek for copper), present as umbers and sulfide stockworks. Podiform Cr-spinel ores in the ultramafic part of the complex are an important source of refractory raw materials.

Papua–New Guinea ophiolite belt. This belt runs the length of the Owen Stanley Ranges in E Papua and consists of several allochthonous units of Palaeogene oceanic crust obducted onto a basement of Mesozoic and Palaeogene sediments and granites. No sheeted dykes are recognized, but the belt includes all the other units of an ideal ophiolite. The Tertiary sediments beneath the ophiolite are deformed and metamorphosed, probably suffering this during the obduction of the ophiolite nappe. What is distinctive here is the continuing activity in the belt, where seismic activity continues on active thrust faults and basement continues to adjust to the isostatic changes consequent to obduction. Active volcanism continues, with magmas rising through the obducted allochthon, presenting the unusual circumstance of an ophiolite obducted onto an active arc during the continued collapse of the Solomon Sea oceanic basin against the Australasian continental foreland (Jaques and Robinson, 1977).

Ballantrae Complex, SW Scotland. This small Ordovician ophiolite is extensively fragmented by later faulting, and is partly buried beneath younger Palaeozoic sediments. All the elements of an ideal ophiolite have been recognized although outcrops of a sheeted dyke complex are very limited. The ophiolite includes extensive trondhjemite intrusions, auto-metamorphic beerbachite dykes and a diverse suite of volcanic rocks, including true ocean-floor basalt, shallow-water hyaloclastic breccia and tuff, island-arc tholeiites, and alkali basalt. The volcanic sequence probably includes a collage of tectonically juxtaposed units originally developed in separate environments. A metamorphic sole complex is present, including glaucophane (blue) schists related to obduction and eclogites related to mantle decoupling. Radiometric ages on the trondhjemites and the blue schists suggest that only ca. 10 Ma (or less) separated generation at a spreading ridge from

obduction (Church and Gayer, 1973; Treloar et al., 1980; Bluck, 1981; Jelínek et al., 1984).

Relationships

The spatial relationships between ophiolites, olistostromes, and tectonic *mélange* are well documented from Newfoundland, Oman, Sierra Nevada, the Alps, Papua–New Guinea, and the Himalayas. Two types of relationship are seen: (1) ophiolite nappes overriding olistostromes that contain both ophiolite olistoliths, and olistoliths of continental shelf sediment, or olistostromes devoid of ophiolite blocks, and (2) tectonic *mélange* that consists of mainly ophiolite fragments in a matrix of pelagic shale and graywacke with blue schists.

The first type possibly records the foundering and tectonic collapse of a continental shelf prior to ophiolite obduction, with ophiolite blocks being contributed from the nose of a rising ophiolite nappe. This nappe ultimately overrides, tectonizes, and thermally alters its own debris, creating a “ductile blanket.” The second type represents a megabreccia accreted onto the trench wall during subduction of oceanic crust, and does not relate to the emplacement of an ophiolite nappe. Metamorphism and deformation here relate to subduction rather than obduction.

Obduction—processes and effects

The oldest ophiolite complexes recognized to date (1986) are late Precambrian (ca. 800 Ma) in age (Fig. 1). If ocean crust has formed and been consumed at rates comparable to those at present, then present ocean crust represents about one-third of all ocean crust that has existed since that date. In contrast, ophiolite complexes account for an infinitesimal fraction of the two-thirds that has been destroyed. The processes that lead to obduction are not the norm for destructive plate margins, and this may imply either that ophiolites represent abnormal ocean crust, or that obduction is not a “normal” destructive margin process.

Spray (1984) provides a review of the evidence for the nature of the obduction process and the “causes” of ophiolite obduction, particularly with regard to the properties of ocean crust that favor obduction rather than subduction. The parameters defined here do not apply to ophiolite fragments preserved in tectonic *mélange* accreted onto the trench wall subduction complex, but to those ophiolites that form all or part of orogenic allochthons.

Age of the ocean crust is of paramount importance. After generation at the mid-ocean spreading ridge, this crust undergoes a maturation process characterized by thermal decay.

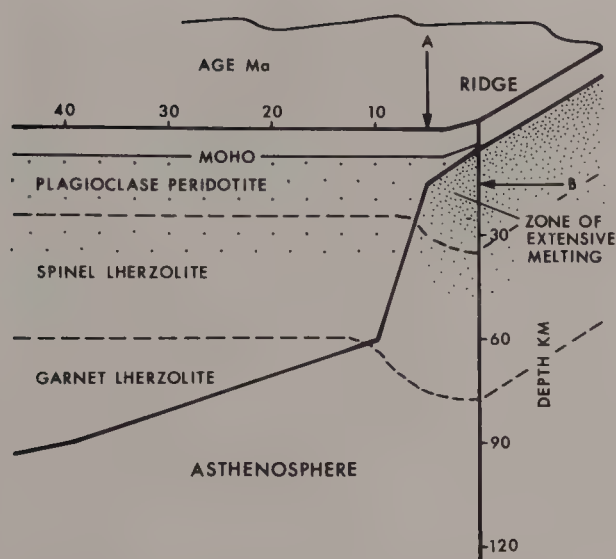


FIGURE 3. Thermal decay of oceanic crust and depth of decoupling shown by profile of oceanic lithosphere overlying asthenosphere; coarse stipple represents depleted mantle (mainly harzburgite-dunite) that overlies relatively undepleted mantle (lherzolite); fine stipple corresponds to area of extensive melting beneath ridge; arrows A and B indicate likely maximum depths and age of lithosphere in order to facilitate ophiolite inception. Horizontal scale is much reduced relative to the vertical. (After Spray, 1984.)

The base of the “instantaneously rigid lithosphere” (that which yields by brittle failure under high strain rates) becomes deeper as the ocean crust ages and cools; from only 2–3 km beneath the ridge crest to something in excess of 90 km for ocean crust over 40 Ma old. This boundary (an evolving transition) effectively marks the top of the asthenosphere (the seismic low-velocity base of the lithosphere) and the most likely site of the decoupling that precedes obduction (Fig. 3).

Density of the ocean crust increases as the crust cools, reflected in the bathymetry of the ocean floor away from the spreading ridge axis, and there comes a point where the thermal maturation produces a cold, dense, thick, rigid slab that from density considerations is more likely to be subducted than obducted. Spray considers 5 Ma old, 15 km thick, rigid lithosphere to be optimal, i.e., most likely to be obducted.

These considerations lead to the following conclusions. (1) Ophiolites contain crust that was only 0–10 Ma old at the time of obduction. (2) Obduction of ocean crust will only occur where a mid-ocean spreading ridge lies within a few hundred kilometers of a destructive margin (e.g., the E Pacific Rise off western N America, or a spreading center that lies in a back-arc basin, or other marginal setting). (3) Where

“layer 4” is preserved in ophiolites it should be dominated by mantle peridotites with the plagioclase lherzolite mineral facies, with possible subordinate spinel lherzolite. (4) Mechanical considerations require maximum temperature about the decoupling zone to be no greater than 1,000°C.

The following evidence supports these conclusions.

1. *P–T* estimates from basal tectonite zones in “layer 4” material fall in the range 800–1,100°C at between 4 and 15 kb. This is consistent with depths between 12 and 40 km. The presence of mafic segregations (pyroxenites, particularly) suggests the presence of a partially melted mantle assemblage such as may be present in the low-velocity zone.

2. Radiometric age dates from the layer 3a magmatic rocks and the metamorphic sole (i.e., reflecting ridge formation and obduction events) commonly fall within 10 Ma of each other, or have overlapping errors, implying it is young ocean crust that tends to be obducted.

3. The sediments and volcanic rocks of layer 1 frequently show a heterogeneity consistent with an arc, intra-arc, back-arc or marginal basin environment for deposition or eruption.

4. *P–T* estimates derived from obduction aureoles, and the general metamorphic facies seen, imply a heat source that is a combination

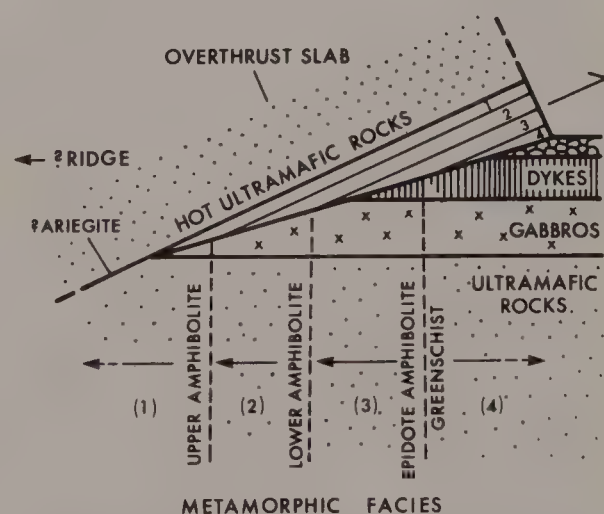


FIGURE 4. Simplified model for the formation of a metamorphic sole along a thrust fault penetrating oceanic crust and upper mantle. The sole is shown divided into four arbitrary “*P–T* planes” ranging from greenschist to upper amphibolite–granulite facies. The sole is only complete when all four *P–T* planes are tectonically juxtaposed. This obviates the necessity of having extremely high thermal gradients existing through the whole thickness of the complete sole. As thrusting and metamorphism continue, lower-grade rocks are progressively “welded” onto the hanging wall such that the locus of shearing moves downwards and away from the peridotite. (After Spray, 1984.)

of residual heat in the obducted slab, and shear heat generated during obduction. Residual heat, equivalent to a sole T in excess of 900°C is essential if the shear heat component is to raise temperatures to those consistent with the granulite and upper amphibolite facies assemblages present in some obduction aureoles (Fig. 4).

Intermediate crust

Some "ophiolite complexes" do not readily fit into the pattern of an "ideal ophiolite." These include such well-known ultramafic massifs as Cerro Ronda in SE Spain and Étang du l'Herz in the Pyrenees. These enormous ultramafic massifs consist predominantly of mantle peridotites, including plagioclase-, spinel-, and garnet-facies assemblages, with high-temperature, high-pressure metamorphic soles, but lacking any well-defined remnants of ocean layers 1, 2, and 3. Other "ophiolites" carry remnants of layer 1 but with little or no volcanic components, an abundance of terrigenous sediment, and occasionally extensive basalt sill complexes. Nicolas (1985) has proposed the existence of an intermediate "metasedimentary" crust to account for such assemblages that formed in spreading basins like the southern Red Sea, the Gulf of California, and the Salton Sea trough. In such situations, the stretched continental crust of a pull-apart basin is breached and a high sediment input fills a basin floored by a rising mantle diapir. Melting in the mantle diapir generates magmas that rarely, if ever, erupt but are largely frozen into the sediments as sills. In this way, typical layers 1, 2, and 3 would not develop as separate entities. Water penetrates the mantle diapir, leading to extensive serpentinization of the ultramafic rock. Heat flow in such situations is extremely high ($\geq 50^{\circ}\text{C}/\text{km}$) and, as a consequence, the base of the sedimentary pile could undergo thermal metamorphism under only 4 km of cover in the epidote amphibolite or even cordierite amphibolite facies.

Economic aspects

The umbers and stockworks of layers 1 and 2 in the Troodos Complex of Cyprus have been worked for more than 2,000 years. The chalcopyrite-pyrite ores, from what is now called "Cyprus-type" stratabound sulfide ores, constitute an important source of Cu, Ag, and Au, and a minor source of Pb and Zn worldwide. Umbers, likewise, are worked for their Cu, Co, Mn, and Ni content, both for the metals themselves and as pigments. Podiform chromite deposits in layers 3 and 4 provide metallurgical grade Cr ore in Cuba, Albania,

and the Philippines, and refractory raw materials in Cyprus, Turkey, and a host of small deposits around the world. On New Caledonia, laterite developed on ultramafic rocks is one of the world's largest Ni resources, and other laterites in the Philippines and Cuba may prove valuable sources of Ni and Cr. The chrysotile asbestos and talc deposits of New England (U.S.A.) and eastern Canada, and scattered through the Caledonides of Norway, are all located in altered ultramafic rocks in ophiolite complexes. These and other ophiolite-related mineral deposits are reviewed by Augustithis (1980) and Hutchison (1983).

ADRIAN F. PARK

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Cross-references: *Alpine mafic magma stem*; *Aureole*; *Geosynclinal volcanism*; *Mid-ocean ridge and ocean-floor petrology*; *Steinmann Trinity*; *Ultramafic rocks*; *Upper mantle—experimental petrology*.

ORBICULAR ROCKS

Orbicular rocks are nonsedimentary rocks that contain concentric shells of different texture and/or mineralogy about a central core. Occurrences have been described from every continent, and orbicules have been found in ultramafic to felsic, igneous, and metamorphic rocks. Host rocks are fine- to coarse-grained, and include dykes, sills, stocks, batholiths, lavas, gneisses, schists, and migmatites. Orbicular facies are generally local, less than several hundred meters in greatest dimension. Other phenomena often associated with orbicules include xenoliths, mineralogical layering, non-orbicular nodules or mineral segregations, phenocrysts, porphyroblasts, and “proto-orbicules.”

Orbicules range from <3 cm to >30 cm in diameter. Shells number from one to more than twenty, and may be spaced irregularly or in geometrical progression. Cores may be fragments of foreign rock (xenoliths), early segregations from a magma (autoliths) or, in metamorphic surroundings, recrystallized country rock, metamorphic differentiates, or remnants of premetamorphic rock. Early-formed orbicules or fragments of orbicules may also serve as cores. In some orbicules, cores are absent (Fig. 1).

Orbicules may be in a matrix distinct from surrounding country rock or occur in otherwise normal facies. They may be closely crowded together, or thinly disseminated. They are usually spheroidal or ellipsoidal, but may be irregular or fragmented owing to deformation, assimilation, or replacement. At a given locality, all orbicules may be similar, or several distinct types may occur. In some localities, there are series of orbicules ranging from single- to multi-shelled types.

Textures of cores and shells may be granular,

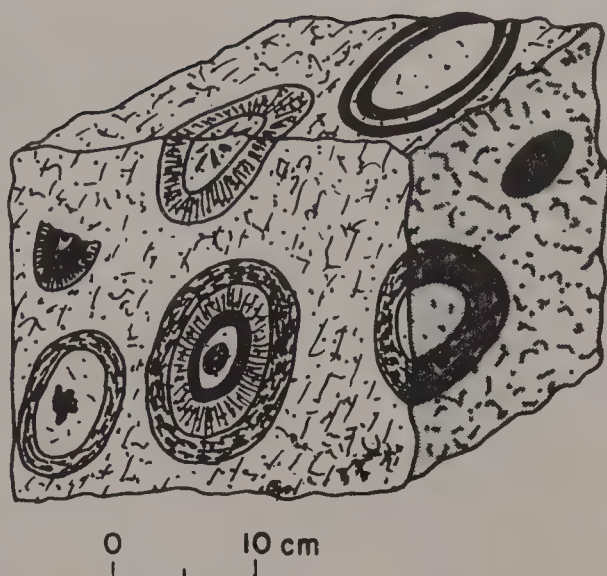


FIGURE 1. Orbicular rock containing whole and fragmented, single and multi-shelled orbicules that display radial and tangential textures.

or have radial and/or tangential orientations owing to the presence of aligned feldspars, amphiboles, pyroxenes, or micas. Mineralogically, orbicules may be similar to or distinct from the country rock.

The heterogeneity of orbicules implies that no single explanation of origin is applicable to all occurrences. Theories of genesis of orbicules by magmatic processes invoke (1) immiscible drops in magmas, (2) reaction between magma and inclusion of foreign rock, (3) fluctuations of a melt about a eutectoid produced by variations in temperature and pressure, (4) migration of a core through portions of magma of varying composition, and (5) rhythmic supersaturation and crystallization about centers in a magma in a manner analogous to the formation of Liesegang rings in a gel. (6) A theory invoking metamorphic processes suggests that the orbicules were formed during granitization by a process of diffusion and rhythmic precipitation similar to that operative in the formation of Liesegang rings.

Reviews of the nature and genesis of orbicular rocks are given by Leveson (1966) and Elliston (1984), with references to earlier works cited therein.

The following terms have been used in relation to orbicular rocks.

Allothraumatic—the nuclei of the orbicules consist of xenoliths differing in composition from the groundmass.

Crystallothraumatic—early-formed phenocrysts form the nuclei of the orbicules.

Esboite—an orbicular diorite, formed by esboitic crystallization in which andesine or

oligoclase is the dominant plagioclase and forms the orbicules (Esbo, Finland).

Heterothraumatic—various kinds of rock or mineral fragments form the nuclei of the orbicules.

Homeothraumatic—the nuclei of the orbicules are formed of inclusions of the same generation as the groundmass.

Isothraumatic—the nuclei of the orbicules are formed of the same rock as the groundmass.

Rhythmic crystallization—the formation of orbicular texture as the result of different

minerals crystallizing in concentric layers.

D. J. LEVESON

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Cross-references: *Enclave*; *Granite*; *Rapakivi texture*; *Xenolith*.

P

PACIFIC AND ATLANTIC SUITES

The concept of a "*Pacific suite*" of igneous rocks as opposed to an "*Atlantic suite*" was developed by Alfred Harker and his contemporaries around the turn of the century. The distinction depends upon the allied concepts of "consanguinity" among rocks and of petrographic provinces. With time, the perception grew that the two supposedly contrasting suites were probably end-members of a nearly continuous spectrum of gradation.

According to Iddings (1892), consanguinity among a group of rocks is expressed by some community of chemical, mineralogical, and textural characteristics, shared by the whole group, either in common or displaying serial variation (cf., Tyrrell, 1955). This is taken to indicate some community of origin, and formed the basis for a genetic classification.

The concept of petrographic provinces, introduced by Judd (1886) and subsequently refined by many workers, is in a sense the embodiment of consanguinity. A petrographic province is a tract of country within which the igneous rocks erupted or emplaced within a given period of time have a certain consanguinity, which may include characteristics that are consistently the same or show gradual changes.

Study of many such provinces demonstrated that many show similar characteristics while being widely separated geographically and in time. Larger scale groupings were thus developed, for instance the concept of "kindreds" (Tyrrell, 1955). Iddings (1892) recognized an "alkali group" and a "subalkali group," to which Harker (1909) gave the names "Atlantic branch" and "Pacific branch," in the opinion that all igneous rocks can be defined as lying along two main lines of variation.

The following are the principal characters of the Pacific and Atlantic suites of Harker.

1. Comparing types of like silica content, (a) the rocks of the Pacific suite are relatively higher in CaO and lower in alkalis; hence, (b) the Pacific suite is serially calcic and the Atlantic suite is serially alkalic, as reflected by (c) relatively more Na-Ca plagioclase feldspars in the rocks of the Pacific suite as

opposed to relatively more alkali feldspars in the rocks of the Atlantic suite.

2. Comparing types of like color index, the rocks of the Pacific suite are more silica-rich, and hence contain more quartz than Atlantic suite rocks, some of which contain feldspathoids, which are unknown in rocks of the Pacific suite.
3. The Pacific suite is best typified by the assemblage basalt-andesite-dacite-rhyolite and the Atlantic suite, by the assemblage olivine basalt-mugearite-trachyandesite-trachyte.

The names Atlantic and Pacific branches came from Harker's contention that the Cenozoic volcanic rocks of the Earth divide into two great petrographic regions, each with its subordinate individual provinces. These two regions of "alkalic" and "subalkalic" rocks corresponded generally with the areas of the Atlantic and Pacific types of coast as defined by E. Suess. These two types of coast are effectively "orogenic" and "nonorogenic," and the two branches of igneous rocks were therefore related to tectonic environment.

The discovery of "Atlantic suite" rocks on the islands in the Pacific Ocean, and advances in understanding the complexity of igneous petrogenesis, blurred the distinction between the two suites. The terms gradually fell into disuse through the early decades of this century. Peacock (1931) devised an "alkali-lime index" as a point on a continuous scale of variation, in part to illustrate the lack of bifurcation in igneous rock composition. Although now recognized as end-members of a more or less continuous spectrum, provinces of "alkalic" (Atlantic) and "calc-alkalic" (Pacific) rocks could still be identified, and this subdivision seemed to be reflected in their distribution between areas of block-faulting and orogenesis. Particularly, the distinction could still be made between the rocks of the circum-Pacific island arcs and mountain belts, and those on ocean islands and in continental rift valleys.

Following the Second World War, sampling of the deep ocean floor revealed the existence of a new major petrographic suite (Engel et al., 1965). Igneous rocks from the ocean floor proved to be quite unlike those on ocean

islands, having lower alkali content, higher silica content, and, particularly, much less potassium. This tholeiitic suite was particularly associated with volcanic activity on the mid-ocean ridges.

The advent of the plate tectonic theory provided a rationale for the distribution of Atlantic and Pacific suites, but also showed that alkalic rocks alone are not the only product of crustal rifting, nor do they always indicate extensional tectonic settings.

Green (1971), using insights from experimental petrology, argued that the magmas erupted in extensional belts and rift valleys varied in alkalinity according to the depth at which magma separated from a crystalline residuum in the mantle source region. Along the mid-ocean rises, where newly formed lithosphere is still hot and thin, segregation of the magmas occurs at relatively shallow depths, and the subalkalic ocean tholeiites result. In rift valleys, and beneath ocean islands where rifting, and fracturing is only incipient, this segregation occurs at greater depths, and involved the melting of a smaller fraction of the mantle source region. More alkalic magmas result.

The calc-alkalic rocks of island arcs and the circum-Pacific mountain belts are associated with inclined zones of seismicity marking the point at which ocean crust is returned to the mantle. The complexity of igneous rocks in the areas above these subduction zones is explained by the synergistic effects of melting of the subducted ocean crust, partial melting of overlying mantle and deep continental crust, and the contaminating effects of water, subducted pelagic sediment, and overlying continental crust.

The concept of worldwide, tectonically influenced associations of igneous rocks has survived, and Harker's distinction between an orogenic "Pacific" and anorogenic "Atlantic" suite, has likewise survived, with a major modification (Martin and Piwinski, 1972) namely, the recognition of a third grand grouping on the ocean floor.

W. R. DICKINSON

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Cross-references: *Alkaline rocks—undersaturated; Basalt; Batholith; Calc-alkaline rocks; Igneous rocks—tectonic setting; Magma; Ocean-island igneous rocks; Rift valley volcanism; Volcanoes and volcanism.*

PAIRED METAMORPHIC BELTS

Parallel belts of metamorphic rocks of similar age but of contrasting metamorphic mineralogy were first recognized in Japan (Miyashiro, 1961) and subsequently this has been recognized as a concept of major importance in global tectonism, since these belts indicate the site of subduction zones at sialic crustal margins. The island of Honshu is composed of two pairs of such metamorphic belts; in each case the inner belt (on the continental side) is a low-pressure metamorphic belt characterized by andalusite–sillimanite or Abukuma-type facies series, while the outer belt next to the oceanic crust is a high-pressure belt of jadeite–glaucofanite or Alpine-type facies series (Fig. 1). The recognition that such metamorphic belts are closely connected with plate margins and that subduction zones lie beneath high-pressure metamorphic belts led to the recognition that the paired belts represent the sites of ancient subduction zone margins, and are diagnostic of the cordilleran type of orogenic belt (Dewey and Bird, 1970).

The high-pressure belts are composed of thick graywackes and turbidites deposited in an oceanic trench over a subduction zone. Frequent high-magnitude earthquakes cause the slumping of large masses of rocks from the shelf and the ocean floor side into the trench to form "wildflysch" type deposits. The oceanic crust of basic pillow lavas and dykes is the "basement" for this sequence and this crust is carried down into the mantle on top of the descending slab dragging with it the thick sedimentary pile (Fig. 2). Subduction zones are areas of very low heat flow, possibly because they are zones of downgoing mantle convection or simply owing to the cooling effect of the large, cold descending slab. Because of this, the sediments and sea-floor volcanic rocks are taken down to



FIGURE 1. Three pairs of regional metamorphic belts in Japan. (After Miyashiro, 1973.)

much greater depths than is normal without a greatly increased temperature. Thus, the rocks recrystallize in the stability fields of lawsonite and jadeite-quartz, giving rise to metamorphism at glaucophane-lawsonite schist facies. The oceanic crust of basic rocks is raised to eclogite facies of a particular type associated with glaucophane schists. In some localities, pillow lavas are found with eclogite facies mineralogy, indicating that the transport to

great depth takes place without strong deformation (Bearth, 1959).

As the descending slab goes farther down, the hydrous basic volcanic rocks reach a depth and temperature at which partial melting may take place. Andesitic magma is generated, varying in composition according to the depth at which the melting takes place. The uprise of these magmas may cause melting in the mantle above the descending slab and also melting and mobilization of the sialic crust in the area behind the oceanic trenches. Thus, the second belt of metamorphism, also caused indirectly by subduction, is characterized initially by chains of andesitic volcanoes with subsequent intrusion of high-level granitic plutons into a zone of rocks that has been regionally metamorphosed by an areal uprise in geotherms well above normal. The characteristic sequence of metamorphic zones is marked by the presence of andalusite and cordierite at lower temperatures and by sillimanite at higher temperatures.

Mechanisms by which such paired metamorphic belts are uplifted and brought to the surface are not as clearly understood as the mechanism of formation, which is essentially during subduction. In different cases, the uplift is due to different causes, although active subduction along the original line must cease. The resultant pair of metamorphic belts do, however, show many other characteristics besides their metamorphic facies series that enable them to be clearly recognizable in older orogenic chains. The outer, high-pressure belt is normally composed of flysch turbidites, and slump deposits. Volcanic rocks are basic pillow lavas, probably all brought in to the area as the oceanic crust. Tectonism and obduction emplaces great blocks of mantle either as serpentinite lenses or as large peridotite bodies with a distinctive mantle fabric, and some may even have oceanic crust, dyke swarms, and metamorphosed gabbros associated with them. Eclogite facies basic masses may be similarly emplaced. Generally, deformation is not as intense as in collision-type orogenies and there is a distinctive and notable lack of granitic activity.

The low-pressure belts are characterized by platform-type sediments with acidic and intermediate volcanic activity at the same time as the sedimentation. Later uprise of geotherms to high temperatures without a concomitant rise in pressure is followed by the widespread intrusion of high-level granites. Frequently, deformation here is in the form of gravity-induced structures from the uprise of the thermal dome of sialic material (Fig. 3).

Typical paired belts are now recognized all around the Pacific Ocean of Mesozoic and Cenozoic age. Their occurrence in older

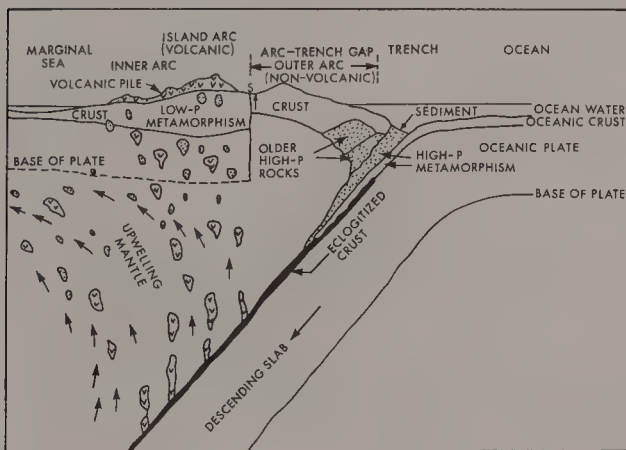


FIGURE 2. Schematic representation of the origin of paired metamorphic belts with special reference to Japan. (After Miyashiro, 1973.)

orogenic chains in other areas is at present in doubt, although there is support for such an interpretation of the Hercynian orogeny in Europe.

The characteristic of collision-type orogenies

(Dewey and Bird, 1970) is that the sediments at the margins of the two sialic plates may be taken down to great depths, but under a normal temperature regime so that high-temperature-high-pressure metamorphism of kyanite-

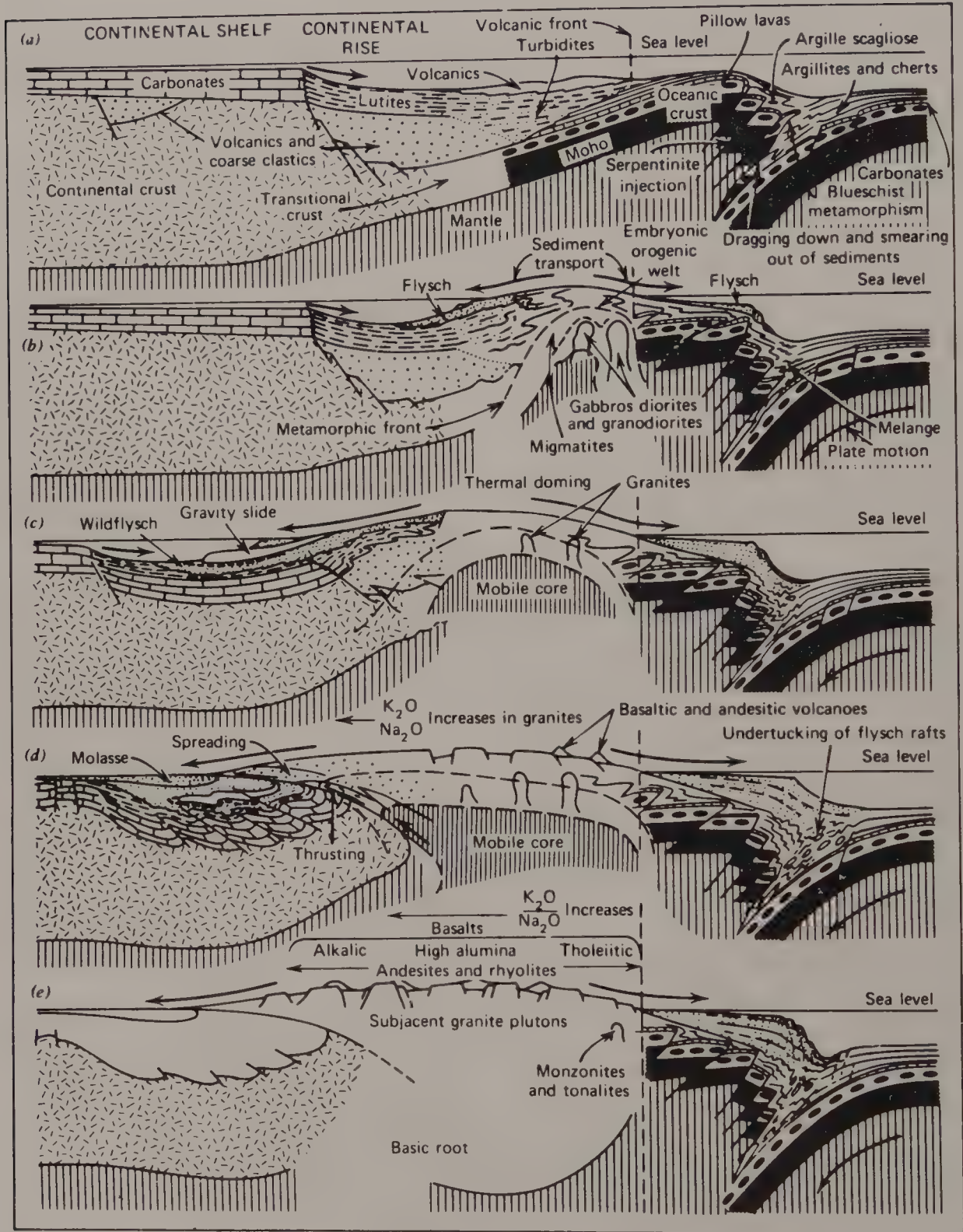


FIGURE 3. Sequence of schematic sections illustrating a model for the evolution of a cordilleran-type mountain belt with the development of paired metamorphic belts by underthrusting of a continent by an ocean plate. (From Dewey and Bird, 1970.)

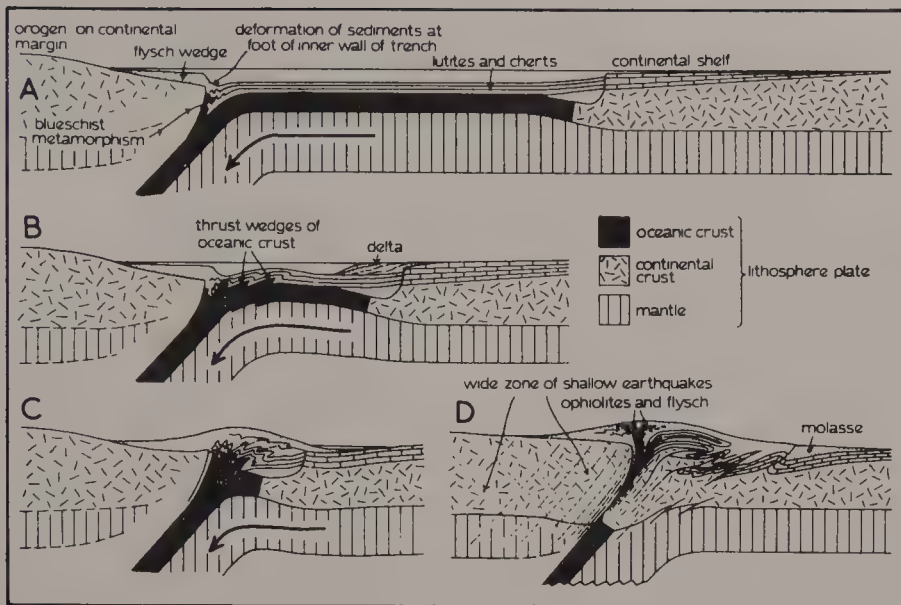


FIGURE 4. Schematic sequence of events forming a collision-type orogenic belt; in D kyanite-sillimanite facies series metamorphic rocks develop in the sediments and slices of basement of the downgoing slab and this is followed by the uprise of granitic plutons. (After Dewey and Bird, 1970.)

sillimanite or Barrovian-type facies series occurs. Slices of ophiolites with glaucophane-lawsonite schists may be obducted (carried up on to the advancing sialic plate) and thus preserved as a parallel high-pressure, low-temperature belt. In this case the two belts are in close tectonic contact, and the high-pressure, low-temperature belt may be sparsely preserved (Fig. 4).

The concept of paired metamorphic belts proved very popular in the late 1960s and 1970s, on the basis of data from the circum-Pacific orogenic belts. As palaeomagnetic data became available on a more widespread, detailed, and reliable basis it was realized that many paired belts had not originated in their present configuration, and the large faults often seen separating members of a pair represent planes of suturing, or massive strike-slip displacement (especially in western N America). The concept of accretion tectonics, in which allochthonous terranes play a major part, has been advanced to explain the microplate collage seen in the NE and W Pacific belts. These concepts are summarized by Hashimoto and Uyeda (1983), who include a reinterpretation of the paired metamorphic belts of Japan.

A. E. WRIGHT

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Cross-references: *Circum-Pacific plutonic and volcanic activity*; *Eclogite*; *Glaucophane schist facies*; *Igneous rocks—tectonic setting*; *Metamorphic facies series*; *Ophiolite*.

PALAEOVOLCANOLOGY

Palaeovolcanology is the study of ancient volcanoes and their products. It has been one of the pillars of uniformitarianism in geology, ever since J. Guettard in 1752 recognized the extinct volcanoes of the Auvergne by their shape and N. Desmarest, eleven years later, demonstrated the identity of basalt as solidified lava. This study provides our only insight into the internal structure of volcanoes and their roots; it also reveals something about the nature of volcanicity in inaccessible environments like deep basins of deposition. As a subdiscipline it is seldom formally distinguished from volcanology because the designation of a suitable dividing

line between "past" and "present" is not a simple matter. It is often impossible to decide whether a volcano is dormant or extinct and upon viewing the Earth as a whole, volcanic activity appears to have been essentially continuous. The International Association of Volcanology and Chemistry of the Earth's Interior (IAVCEI) approved the publication of a map showing the post-Miocene volcanoes of the world; this excludes all centers of eruption that have not been active during the last 7 Ma and that are likely to have been considerably denuded. On the whole, the end of the Miocene appears to be a reasonable limit but it remains an arbitrary choice; e.g., in the Tasmanian province (Sutherland, 1969) volcanism continued from Eocene times at least into the Pliocene and perhaps into the Pleistocene (i.e., from ca. 45–1 Ma ago).

If volcanism is a normal component of the geologic processes that are continuously modifying the Earth, lavas and pyroclastic deposits may be expected to be at least as prominent among the oldest as among the youngest rocks. This is undoubtedly so, but significant differences in time have been found as far as distribution, type of volcanic activity, amount of erupted material and perhaps also chemical composition of lavas are concerned. Whereas contemporary volcanoes, for instance, are restricted to certain well-defined tracts of the Earth's surface, no large part of the crust has entirely escaped volcanic activity in the past. It seems most likely that the basic structural conditions leading to volcanism have remained the same but that these were attained at different localities at different times. These and other considerations have led to a classification of volcanic eruptions according to the geotectonic environment in which they occur. Thus, certain kinds of activity are typically oceanic, other types are characteristic of the stable, continental platforms (cratons) and still others are associated with the mobile belts (orogens) in which geosynclinal deposition of sediments is followed by mountain building movements.

In terms of plate tectonic theory, the first category would include constructive plate boundaries (i.e., mid-ocean ridges) as well as oceanic intraplate volcanism. The second category may be equated with continental intraplate activity, and the third with destructive plate boundaries (i.e., subduction zones). Attempts have been made to recognize lavas characteristic of different plate tectonic environments by their geochemical signatures even in the oldest rocks. However, the extent to which plate tectonic concepts are applicable to the earlier part of the Earth's history is still being debated. In the following paragraphs, modern

settings are first described with respect to present plate boundaries. Similar but older volcanic rocks are then interpreted as having formed in ancient settings analogous to the modern ones.

Oceanic environment

The deep ocean basins are floored by great thicknesses of basaltic lava. This was predicted from seismic data and confirmed by dredging and drilling. The lavas are thought to have welled up quietly through fissures like those in Iceland along the crest of the Mid-Atlantic Ridge. Underwater photographs have shown that submarine fissure eruptions (as may be expected) did not spread out laterally but were rapidly chilled and piled up in pillow-like shapes. According to modern ideas about the evolution of the Earth, the sites of these fissures are determined by convection currents that rise within the mantle and move away in opposite directions below the crust, creating tension along the mid-ocean ridge above them. New fissures are continually created as the older crust is carried along by the currents. This is called sea-floor spreading.

Virtually all samples of lava from the ocean floors belong to a subalkaline variety of basalt called tholeiite, but abyssal tholeiite differs from other tholeiites in being exceptionally poor in K (<0.2% K_2O) and in the dispersed trace elements Rb, Cs, Sr, Ba, Th, U, Ti, Zr, and the light rare-earth elements. Close relatives of the so-called low-K tholeiites are also found on oceanic islands and in the volcanic arc (i.e., present-day orogenic) environment. These geochemical characteristics of comparatively recent lavas are best explained by tapping at shallow depths (<50 km) of a "residual" mantle from which most of the dispersed elements were previously extracted and concentrated in continental crust before the opening up of present-day oceans. Hart et al. (1970) postulate that this depletion took place since the Archaean because Archaean basalts (though erupted along trenches rather than abyssal depths or mid-ocean ridges) are more like modern alkali basalts than tholeiites in having "nondepleted" trace-element ratios.

In addition, the ocean floors are dotted with single volcanoes forming seamounts (guyots) and oceanic islands built almost entirely of lava. Quiet effusions are preponderant over violent eruptions on the oceanic islands. Only a very small proportion are still active (3% of the world's active volcanoes are located in the Pacific region), affording but a feeble indication of the stupendous amount of activity in the past. Oceanic volcanoes of exceptional size like those

of Hawaii rise 10,000 m above the sea floor and may be associated with archipelagic aprons containing 40,000–50,000 km³ of lava. Most probably the major parts of these volcanoes consist of early tholeiitic basalt, but during the later stages the erupting magma invariably becomes more alkaline and eventually small amounts of mugearite, trachyte, phonolite, and rhyolite may be produced, often associated with explosive activity. This has long been recognized as the typical oceanic alkaline olivine basalt association. Compared with tholeiite, all these rocks are notably high in K + Na, lower in Mg + Fe and have a higher Fe/Mg ratio and Si content; alkaline olivine basalt is also preferentially enriched in the light rare-earth elements. It may originate from mantle peridotite at greater depths than tholeiite by a relatively small degree of partial melting, and gives rise to the later magmas by fractional crystallization.

It has been estimated (Macdonald, 1972) that the rate of oceanic volcanism exceeded that of continental volcanism throughout geological time at least 100-fold. Yet only a small proportion of this stupendous volume of lava has been preserved in the geological column. Geochronology and palaeomagnetism have shown that no part of the ocean floor is older than 200 Ma. Ocean basins that must have existed in earlier times have been destroyed by plate tectonic movement, most of the older sea floor being overridden by continents and returned to the mantle along subduction zones. Oceanic basalt that reaches the continental crust in the orogenic environment does so (1) by obduction as part of an ophiolite suite or (2) after remelting and contamination in the form of calc-alkaline volcanoes.

Cratonic environment

Volcanic activity in stable continental areas is associated with tension due to up-arching, as evidenced by rift valleys, and with simple fracturing. The typical manifestations of intra-cratonic palaeovolcanicity are plateau lavas and layered intrusions, alkaline complexes and carbonatites, diatremes, and kimberlite pipes.

Murray (1970) drew attention to differences in composition and volume between the products of oceanic and cratonic rift valley volcanism and ascribed them to fundamentally different mantle conditions in the two environments: high heat flow under the mid-oceanic ridges and low heat flow under the continents.

Only one cratonic area is active at present, viz., central Africa and the adjacent Arabian peninsula, where 22 active volcanoes are known (Macdonald, 1972). Many more have erupted since the beginning of the Quaternary period.

Most are of a distinctive type: great cones of alkaline olivine basalt, trachyte, phonolite, nephelinite, sodic rhyolite, or rare K-rich lavas and pyroclastic deposits. Explosive eruptions played a prominent part during their development and may be ascribed to a high volatile content of the magma (especially CO₂). Alkali and volatile enrichment coupled with silica deficiency has been carried to an extreme in the carbonatitic volcanoes of which Oldoinyo Lengai is a modern example. The proportion of phonolite and trachyte to basaltic lavas in the E African rift valleys is as high as 1:1 or 1:2. This has led some authors to suggest that the highly alkaline lavas arose in the mantle and were not derived by fractionation from alkaline olivine basalt.

During the geological past, a far greater variety of volcanic phenomena characterized the cratonic environment, some of which are imperfectly understood because modern equivalents are lacking. An excellent example of this association is provided by the Rhine valley and environs in W Germany (Fig. 1).

Lava plains and plateaus. These are built of numerous subhorizontal flows that followed each other in vertical succession and flooded the continents over vast areas (see **Flood basalt**). Except locally, the rate of extrusion was not particularly rapid, for example, about one flow per 50,000 years on the Columbia River plateau, western U.S.A. (Gibson, 1969), while the accumulation of basalt flows in the Parana basin, S America, took no less than 28 Ma (Amaral et al., 1966). Where the lavas have been partly eroded, as in Siberia, U.S.S.R., and S Africa, underlying strata are often riddled with sills and dykes of dolerite or quartz dolerite that crystallized from the same magma. In some provinces (Tasmania, Australia) most of the magma probably failed to reach the surface. Although transitions between dykes and flows are seldom observed, the flows are generally considered to be the result of fissure eruptions that were fed along dykes. Another mode of emplacement that has been invoked is that of broad, overlapping cones with a very low (<1°) angle of slope (Macdonald, 1972).

By far the largest fissure eruption in historic time was the Laki flow of 1783 in Iceland (in a mid-oceanic ridge context). It covered 565 km² and had an estimated volume of 12–15 km³, but this is puny in comparison with the great intracontinental eruptions of the past. The Roza flow in the Columbia River plateau covers at least 25,600 (perhaps 51,000) km² and its volume is of the order of 2,000 km³. The average thickness of individual plateau basalt flows is from 15 m to 30 m and their volumes probably range from 200 to 300 km³.

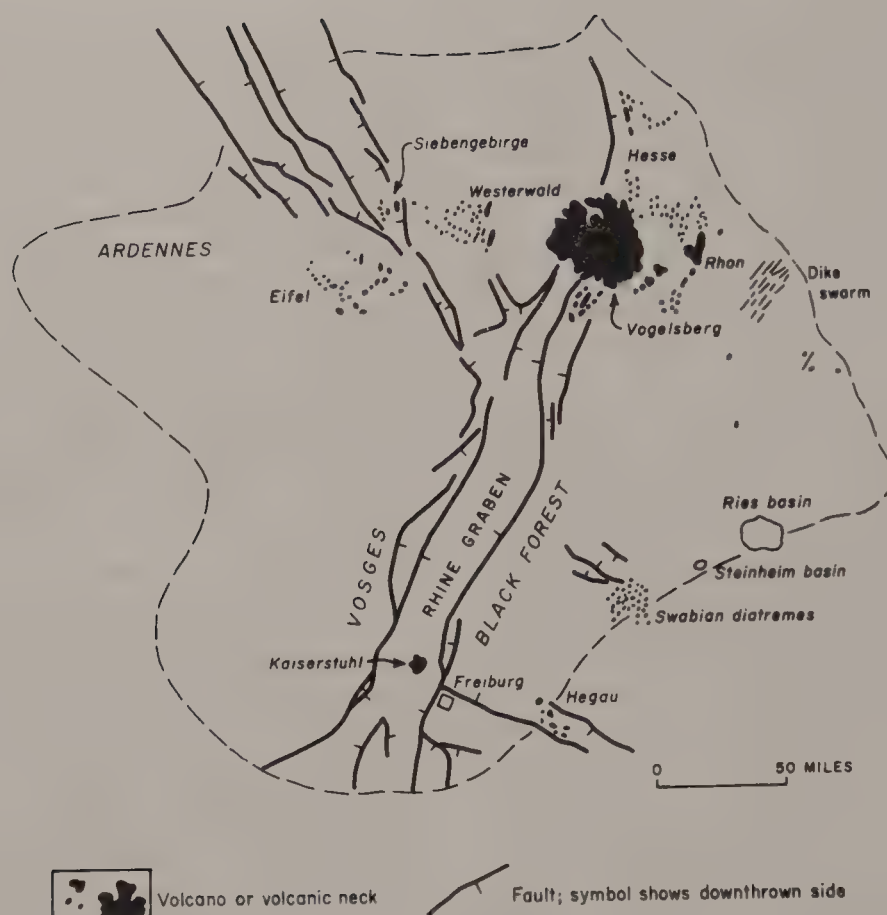


FIGURE 1. Sketch map of the Rhine upwarp showing the association of rift-valley faults and volcanic features of cratonic type. (From Macdonald, 1972.)

Basaltic lava is apparently the only type with sufficient fluidity to form such extensive flows, although Precambrian andesite of the Venterdorp Group in S Africa covered at least 250,000 km² and appears to be analogous to the plateau basalts in its nonorogenic setting. Plateau basalts are typically tholeiitic, although relatively large alkaline olivine basalt flows are known in Ethiopia and elsewhere. The Thulean (N Atlantic) province is exceptional in that it consists of alternating tholeiitic and alkaline basaltic flows. Once thought to have covered 1,800,000 km² between Greenland and Ireland, this province was probably of moderate size before it was disrupted by continental drift.

Apart from their enormous volume, the most striking feature of the tholeiitic flood basalts is their vertical and lateral uniformity in composition and their similarity all over the world; however, minor geochemical differences have been recognized (cf. Rhodes and Krohn, 1972). Continental tholeiite is generally similar to oceanic tholeiite, with low ⁸⁷Sr/⁸⁶Sr, Rb/Sr, Rb/K, Th/K and U/K ratios, indicative of a mantle source, but Tasmanian dolerite is characterized by values more appropriate to a

crustal origin. A temporary rise of the geothermal gradients is required to explain tholeiitic eruptions on the continents.

The most extensive plateau basalts are all Triassic or younger. From a survey of the literature, Ahmad (1972) concluded that another important series of plateau eruptions took place approximately 1,000 Ma ago. Doubts have been expressed, however, whether the Keweenawan lavas of N America, with their great thickness and apparently limited extent, really fit in to the concept of plateau basalt; other Precambrian examples are even less certain. Almost the only one known in the interval between 1,000 and 200 Ma ago is the Antrim plateau in N Australia. The Mesozoic and Cenozoic flood eruptions of the southern continents all lie on present-day continental margins. The conclusion seems inescapable that these vast outpourings of subaerial basalt must be related to the initiation of continental drift, their chronology reflecting the sequence of disruption. Once the continents have moved far enough apart, further effusions of the magma would have been submarine along the mid-oceanic ridge. Continental tholeiites could

therefore be genetically linked with the oceanic environment and their presence on the continents could be regarded as incidental. However, not all flood basalts (e.g., the Columbia River plateau) are directly related to the world ridge–rift system. This led Gibson (1969) to conclude that they simply occur in regions of crustal tension regardless of how that tension is produced.

Layered intrusions. Thick sills of tholeiitic magma that failed to erupt as plateau basalt sometimes differentiated by crystal fractionation into layered units of picrite, gabbro, peridotite, pyroxenite, chromitite, anorthosite, and diorite, as exemplified by the Skaergaard, Bushveld, and Muskox intrusions. In many cases the plutonic rocks crystallized under cover of an acidic volcanic phase that preceded their emplacement. Although they occur in varied tectonic environments, these complexes were rarely affected by orogeny.

Alkaline ring complexes. Near-circular bodies with a concentric zonal structure built of carbonatite, peridotite, biotite pyroxenite, ijolite, nepheline syenite, and syenite, in ideal sequence from the inside outwards, constitute a well-defined igneous association. Their shape, relatively small size (5–50 km²), vertical contacts, brecciated country rock and internal structure leave little doubt that they represent the plutonic equivalents of the alkaline volcanoes already mentioned. Such a relationship is firmly established for the Napak volcano, Uganda (Fig. 2). Like the granitic ring complexes or cauldron subsidence structures of Scotland, Norway, Nigeria, and New Hampshire, U.S.A., many of the alkaline ones were probably connected with the formation of calderas at the surface. H. Backlund first pointed out in 1932 that alkaline complexes are restricted to “the old stable areas.” In Siberia they are localized near the periphery rather than the center of the platform, e.g., where the Aldan and E Sayan upwarps are marked by rift valleys. Alkaline complexes range from Precambrian to Quaternary in age and are

considered to represent the normal type of magmatism associated with cratonic conditions.

Plugs of alkaline lava. Although any volcano may be eroded down to a volcanic neck, plugs of alkaline basalt, trachyte, phonolite, monchiquite, etc. occurring in swarms such as those of the Siebengebirge in W Germany, the Klinghardt Mountains in SW Africa, and the Hopi buttes in Arizona, U.S.A., are characteristic of past cratonic volcanism.

Diatremes. These are volcanic necks consisting predominantly of fragmental material (tuff, breccia) but often intruded by dykes. Examples are the Swabian tuffsite pipes (W Germany) and the Navajo buttes (U.S.A.); kimberlite pipes were probably emplaced in like manner. H. Cloos demonstrated in 1941 that the Swabian pipes are not merely the result of explosive eruptions, but of fluidization by a suspension of ash particles and lava droplets in gas. This process appears particularly applicable to the alkaline, CO₂-enriched magmas of the cratonic environment, but diatremes are also found elsewhere; sometimes because of contact with groundwater, such phreatic and phreatomagmatic eruptions may be highly explosive owing to the effect of pressurized steam. The Pleistocene “maar” volcanoes of the western Eifel district (W Germany) are considered to be surface expressions of alkaline diatremes.

Kimberlite pipes. No other rock type seems to be more dependent on cratonic conditions than kimberlite. Yet it is not a normal volcanic product, because it appeared at the Earth's surface only at rare intervals and in a few areas in very minor quantities. It is a nonequilibrium assemblage with “hybrid” geochemical characteristics. From numerous investigations of diamonds and their inclusions and of mantle-derived xenoliths in kimberlite, its depth of origin has been estimated at 150–200 km. Most of the S African pipes are Jurassic to Cretaceous in age and the Siberian kimberlites range from Ordovician to Jurassic.

Eruptions of kimberlite have not been witnessed but the internal structure of a

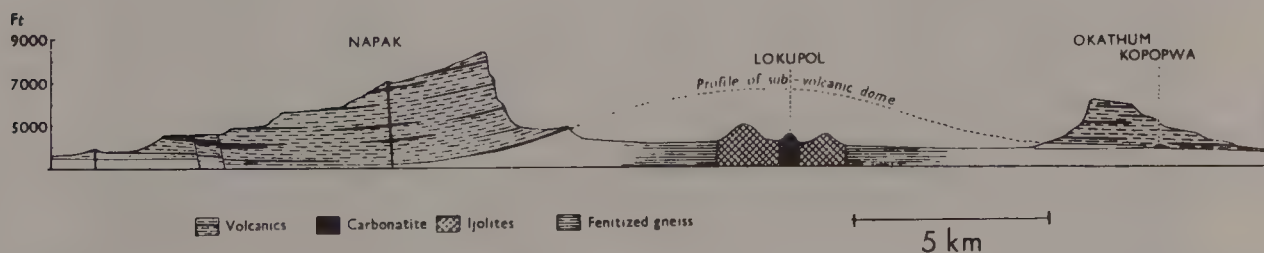


FIGURE 2. Geological section across the Napak volcano, Uganda, with its central alkaline ring complex. (From King, 1965.)

kimberlite volcano is rather well known. Slightly eroded pipes in central Africa, Botswana, and the Bushmanland area of S Africa consist of deep depressions filled with sediments including reworked kimberlite tuff. The pipes of kimberlite breccia underneath the crater deposits contract in depth and eventually merge with kimberlite dykes. The drilling of the vents probably started 2–3 km below the surface.

Orogenic environment

Volcanic rocks are essential components of most great mountain ranges and eroded orogenic belts. While their interpretation is often rendered difficult by folding, dislocation, and metamorphism, these rocks surpass the plateau basalts in volume and provide most of the subject matter in palaeovolcanology.

Orogenic belts are now generally understood in terms of the collision of lithospheric plates, but much of the earlier nomenclature based on geosynclinal concepts of mountain building can be retained because the geosynclinal hypothesis is easily reconciled with plate tectonics.

The following sequence of volcanic episodes corresponding with successive stages of the orogenic cycle has been recognized, although many irregularities occur. During the early depositional phase, submarine basalts are

poured out in a eugeosyncline and become interbedded with sediments. They are mostly tholeiitic pillow lavas (Fig. 3) and hyaloclastites (i.e., debris formed by disintegration of lava in water, glassy when unaltered). Quite common in this environment is a variety of basalt called spilite with all the usual textural features but with albite, chlorite, epidote, and calcite as principal constituents. More silicic lavas and tuffs, which are likewise Na-enriched (keratophyres), are sometimes associated with it. Spilite shows a greater chemical diversity than basalt and was probably more abundant in Palaeozoic and Mesozoic times than before or since. Another member of this association (known as the ophiolite suite) is serpentinite, the origin of which is also debated. At least some serpentinites have been shown to consist of ultrabasic lava flows erupted in very deep water. In recent years they were found to be quite common in the Archaean greenstone belts where bladed skeletal olivine crystals form “bird track,” “spinifex”, or “crystalline quench” textures in peridotite (Viljoen and Viljoen, 1969; McCall and Leishman, 1971).

Many geologists now consider that ophiolites (consisting as they do of ultramafic, gabbroic and basaltic rocks capped by deep-sea sediments) are actually fragments of oceanic plates that have been thrust into a subaerial environment.

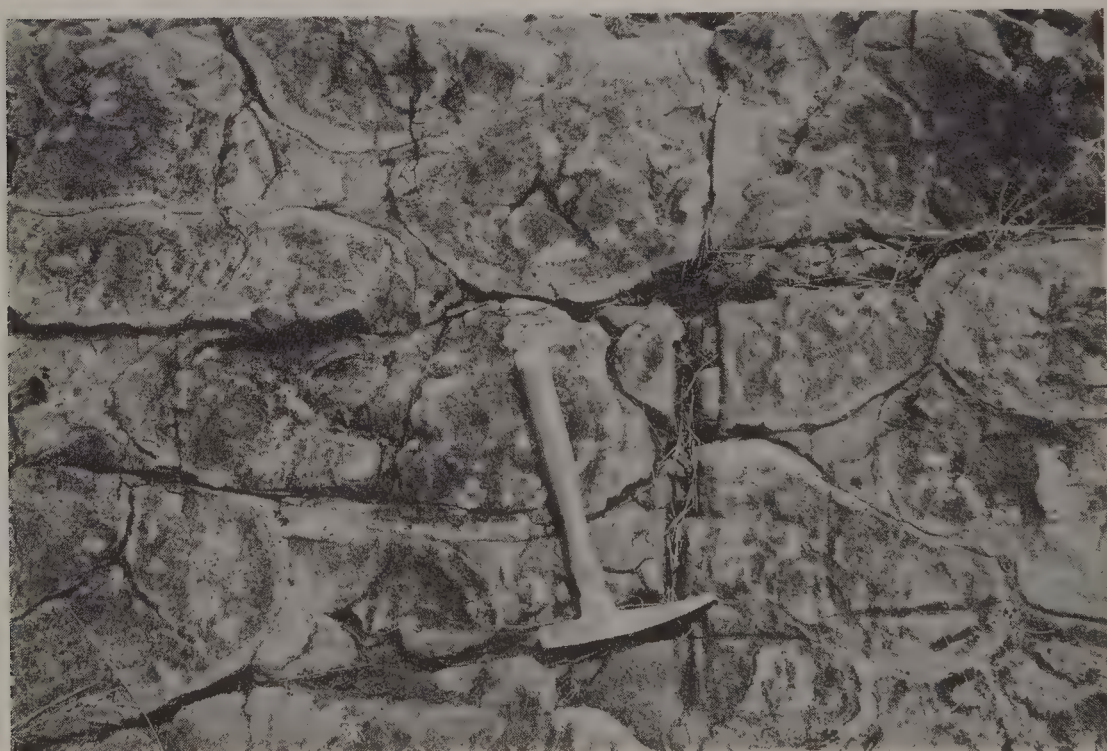


FIGURE 3. Pillow structure in ca. 3.4 Ga old basaltic komatiite of the type area, Barberton, S Africa. (From Viljoen and Viljoen, 1969.)

At a somewhat later stage, a geanticline, or magmatic arc, capped by a row of volcanoes may rise along the edge of the geosyncline above a subduction zone. This contributes ash, mudflows, lava (basalt, andesite, dacite), and sometimes ignimbrite, interlayered or mixed with the sediments. The life of a typical geosyncline is terminated after several hundred million years by folding, thrust-faulting and uplift. During the early stages of folding, volcanic activity is generally lacking. In some cases, the synorogenic emplacement of granitic batholiths, and small subvolcanic stocks of quartz monzonite (adamellite-granite) constituting the economically important porphyry copper deposits, is accompanied by andesitic to dacitic eruptions; this kind of activity is mainly postorogenic and represents the second major stage of volcanism. In the Sierra Nevada of California an interval of 85 Ma elapsed between folding and volcanism, whereas in some parts of the Rocky Mountain chain volcanism started almost immediately (Macdonald, 1972).

The eruptions are typically of the explosive kind due to high viscosity of the magma and they build great composite cones like those of the circum-Pacific belt. The lavas are said to be calc-alkaline (basalt-andesite-dacite-rhyolite) and are also known as the orogenic suite. There is still uncertainty about the derivation of andesite, but earlier views of an essential contribution by continental crust have been superseded by theories based on the underthrusting of oceanic lithosphere plates along Benioff zones. Recent isotopic and trace-element data favor the hypothesis that andesitic magmas are derived largely from the mantle wedge above the subducting slab.

After the composite volcanoes, flood eruptions of tholeiite and high-alumina basalt may take place, as in the Cascade Range, western U.S.A. When the relief of a mountain belt has been considerably reduced by erosion, very large volumes of rhyolitic to dacitic ash and froth flows (ignimbrites) may be erupted. The source vents of many ignimbrite sheets were arcuate fissures around the crest of broad up-arched domes (McBirney, 1970) or caldera faults, as in the case of the Valles Caldera, New Mexico, U.S.A. In other cases (Valley of Ten Thousand Smokes, Katmai, Alaska) the fractures were linear. No large ignimbrite eruption has been observed and the origin of highly siliceous magma in such great bulk is still inadequately explained. Finally, with relief of pressure in an orogenic belt, rift valleys and other tension fractures develop and are often accompanied by volcanism with alkaline affinities, as on the adjacent craton. Potassic lavas like those of the Leucite Hills of

Wyoming, U.S.A., occur outside the fold belt.

Discussion

The question whether lavas arising in the mantle have undergone a chemical evolution in the course of time is usually answered in the negative. Suggestions that magmas have become more alkaline or less sialic may be discounted (Rittmann, 1962). Vallance (1969) emphasized that the original magmatic character of Palaeozoic and older lavas cannot be determined with the same assurance as modern ones. However, Viljoen and Viljoen (1969) concluded from their study of 3.4 Ga metabasalts and metaperidotites in the Barberton area of S Africa that a "primitive" so-called komatiite magma with distinctive geochemical characteristics (low Al, Na, K, and exceptionally high Ca/Al ratio) existed in early Precambrian times and may have originated by almost complete melting of the mantle, before the development of a substantial sialic crust. McIver (1972) has drawn attention to similarities between the ancient basaltic komatiites and hornblendites and appinites of much younger age (1,950 and 450 Ma, respectively). At any rate, it seems certain that conditions were appreciably different in the early Precambrian when depositional troughs were relatively small and alkaline olivine basalt was unknown because the lithosphere was too thin for the requisite pressure and temperature of partial melting.

W. J. VERWOERD

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Cross-references: *Andesite and dacite; Archaean volcanism; Basalt; Diatreme; Flood basalt; Geosynclinal volcanism; Igneous rocks—tectonic setting; Kimberlite; Ophiolite; Ultramafic lavas; Volcanic rocks; Volcanoes and volcanism.*

PALAGONITE

The name *palagonite* was applied by S. von Waltershausen in 1845 to a resinous yellow-brown material, thought to be a new mineral species, occurring in tuff at Palagonia, Sicily. It forms by the low-temperature hydration of the clear basaltic glass (called *sideromelane* by von Waltershausen in 1853) found in hyaloclastites, tuffs, and pillow lavas, which results particularly (but not exclusively) from the quenching of basaltic magma in water (Fig. 1). Volcaniclastic rocks rich in palagonite are widely known as palagonite tuffs or palagonite breccias. Quaternary volcanic assemblages in Iceland consisting largely of such rocks were once called the "Palagonite Formation"; they are now called "Moberg."

Palagonitization, the hydration reaction by which sideromelane changes to palagonite, is approximately isovolumetric, and losses in Si, Mg, Ca, and Na (sometimes also Al) are compensated for by gains in water. FeO is largely oxidized to Fe₂O₃ during the process. The "lost" constituents form authigenic minerals such as zeolites, apophyllite, and CaCO₃ which fill voids in the same rock; the closing of

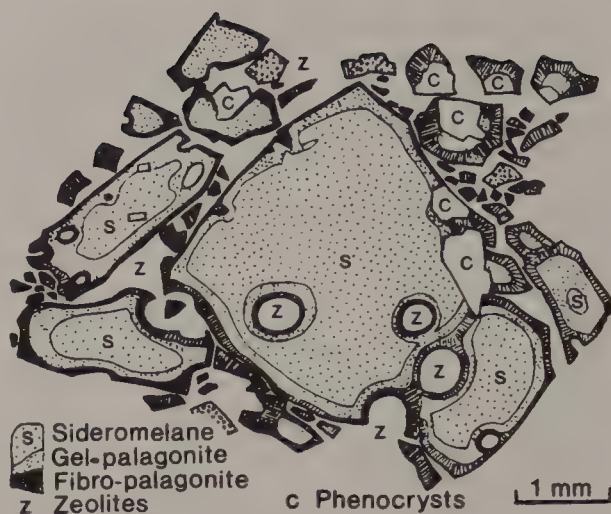


FIGURE 1. Drawing of a thin section of a glassy basaltic breccia (hyaloclastite) from Iceland, showing the relationship between the basaltic glass, sideromelane, and its palagonite alteration product; the clasts are cemented together by authigenic zeolites.

voids by these minerals stops access of water and halts palagonitization. Hence palagonite tends to form a rim around sideromelane bodies. A rim 10 μ m thick forms in 10³–10⁴ years in cold sea water, but formed in 5 years at Surtsey, Iceland, on ash exposed to water at 50–100°C.

Two varieties of palagonite occur. One is optically isotropic, gel-palagonite. It contains 15–30% of water, and typically has a refractive index 0.05 to 0.11 lower than the fresh glass with which it is in sharp contact. The other is microcrystalline fibro-palagonite. It contains less water, has a higher refractive index, and is normally separated from any residual sideromelane by a layer of gel-palagonite. X-ray studies show that it is a multiphase composite substance and is in part a smectite chlorite or montmorillonite. Under low-grade metamorphism it changes readily to chlorite. The term *korite* has also been used for a variety of palagonite.

A secondary greenish or brownish material that is nearly identical with palagonite (although it contains less alumina) is often found infilling amygdalae in basaltic rocks instead of replacing glass in situ and is called *chlorophaeite*.

GEORGE P. L. WALKER

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Cross-references: *Brownstone facies*; *Hydrothermal processes*; *Ocean-floor metamorphism*; *Spilite*; *Zeolite facies*.

PARAGENESIS—See Vol. IVA

PARTITION COEFFICIENTS—See Vol. IVA

PEGMATITE

Pegmatites are holocrystalline usually leucocratic rocks that are, at least in part, exceptionally coarse-grained, and in which extreme variations in grain size are widespread. Their dominant constituents are minerals typically found in ordinary igneous rocks. Some pegmatites also contain less common minerals. Their remarkable features of structure, texture, and fabric most clearly distinguish them from other kinds of igneous rocks. Individual bodies range in maximum dimension from a few centimeters to thousands of meters. Most are tabular in form and 5 m or less in maximum thickness, but included among the remainder are many contrasting shapes and dimensions.

Occurrence

Pegmatites are very widespread but constitute a tiny fraction of the Earth's crust. These distinctive rocks nonetheless have long attracted attention as assemblages of both common and rare minerals with highly varied crystal form, size, and mutual relationships. In many occurrences they have been economically valuable sources of clays, feldspars, gem materials, micas, silica, special fluxes, and

industrial crystals, as well as of Be, Bi, Li, Mo, rare-earth elements, Sn, Ta–Nb, Th, W, and U.

Bodies of pegmatite are most abundant in the shield areas and younger mountain chains of the continents. They are essentially restricted to terranes of igneous and metamorphic rocks; within the prisms and blankets of sediments and sedimentary rocks they occur only as minor associates of scattered igneous masses. Pegmatite bodies of Precambrian age are by far the most numerous and widespread among all those now exposed, owing mainly to the great extent of Precambrian rocks. Pegmatites with diminishing size and frequency occur throughout the geological column; in the Phanerozoic they are largely restricted to the vicinity of batholiths (orogenic and anorogenic) and the high-grade metamorphic core complexes of younger orogenic belts.

Host rocks

The rocks that enclose masses of pegmatite represent nearly the entire span of igneous and metamorphic types. Much pegmatite occurs wholly within domains of metamorphic rocks as bodies formed either by mechanical emplacement, by selective in situ replacement of the country rocks, or by various combinations of processes. Most, however, are closely associated with relatively common kinds of igneous rocks and in numerous occurrences may well be genetically related to them. Thus, many plutons contain pegmatite masses formed by segregation or autoinjection, with examples ranging from small miarolitic pods and stringers to single or multiple enclosures of enormous size. A zonal distribution of pegmatite bodies within and around considerably larger masses of intrusive igneous rocks also has been recognized in many areas, where it commonly is accentuated by systematic variations in size, structure, and composition of the pegmatite bodies.

Structural features of the host rocks appear to have exercised significant control over the shapes and attitudes of most discrete bodies of pegmatite, especially those that are relatively small. Among such features are fractures, faults, breccia zones, bedding and foliation, flow layering, cleavage and schistosity, linear elements, including the axial portions of folds, and boundaries between different rock units. Various contact relationships suggest a complete range from mechanical emplacement of pegmatite-forming fluids accompanied by little or no interaction with the host rocks to pegmatite development by localized melting, recrystallization, or replacement within host-rock terranes.

Many pegmatite bodies are fringed or surrounded by zones of country rock recrystallization or metasomatism ranging in width from less than a centimeter to thousands of meters. Some of these zones are distinguished by concentrations of biotite formed from wall-rock hornblende, chlorite from biotite, or muscovite from kyanite and sillimanite, and others by numerous metacrysts or fine-grained disseminations of minerals such as albite, allanite, apatite, beryl, biotite, diopside, fluorite, garnet, ilmenite, lepidolite, magnetite, microcline, muscovite, topaz, tourmaline, tremolite, and zinnwaldite. These occurrences ordinarily indicate an escape of alkalis, F, and B from the pegmatite systems into the adjacent rocks. Transfer of materials in the reverse direction is suggested by locally abundant andalusite, cordierite, corundum, or sillimanite in some pegmatite bodies enclosed within highly argillaceous metamorphic rocks, by unusual concentrations of calcite, prehnite, and other Ca-rich minerals in many bodies within carbonate rocks, and by marginal concentrations of biotite and magnetite in bodies adjacent to highly ferruginous rocks.

Composition

Most pegmatites are mineralogically simple. Many of them consist of little more than quartz or alkali feldspar, especially in certain extensive gneissic terranes. Nearly all of the others can be likened, in terms of bulk composition, to common igneous rocks in the range granite to gabbro and in the less silicic range syenite to alkali gabbro (Fig. 1). In associations where they appear to be genetically related to such igneous rocks, the pegmatites tend to be somewhat richer in SiO_2 , Na, and K, and poorer in Ca, Fe, and Mg.

As *simple pegmatites* are generally very coarse-grained equivalents of common plutonic igneous rocks, their mineralogy reflects the same compositional spectrum. Thus, although granitic compositions dominated by quartz and alkali or plagioclase feldspars are the most abundant, syenitic, dioritic, gabbroic, and feldspathoidal types are known, corresponding to the QAP classification. Complex rare-mineral pegmatites are more exotic, but these bodies are generally zoned, and some portion usually shows a mineralogical relationship to a common plutonic rock type.

The most common varietal minerals in pegmatites are muscovite and biotite. Amphiboles, calcite, cordierite, fluorite, lepidolite, phlogopite, pyroxenes, spodumene, and black tourmaline (schorl) variously characterize the pegmatites of specific districts or regions.

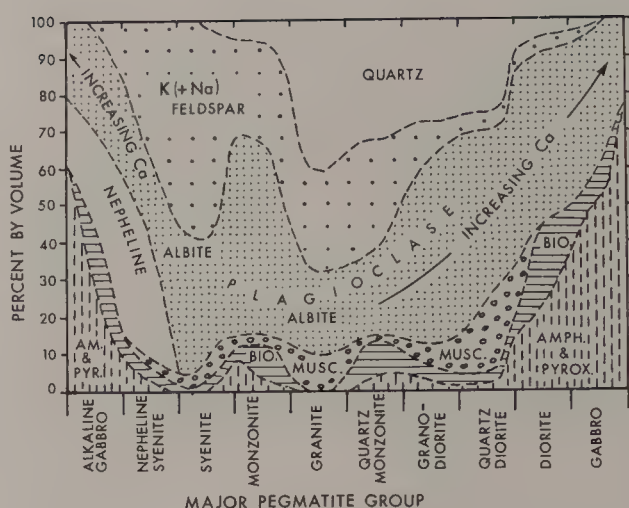


FIGURE 1. General trends in mineral content among pegmatites that are akin to ordinary igneous rocks in terms of gross composition. Only the essential minerals and principal varietal minerals are considered.

Allanite, apatite, beryl, garnet, magnetite, topaz, and zircon are widespread but typically sparse accessory species.

Of special mineralogic interest are the so-called *complex pegmatites*, in which concentrations of relatively rare elements are expressed by the occurrence of uncommon species in exceptional variety, abundance, or crystal size. Much albite of the cleavelandite variety typically is present in these pegmatites, along with one or more minerals containing elements such as Be (beryl, chrysoberyl, gadolinite, phenakite), B (axinite, tourmaline), Cs (alkali beryl, pollucite), Li (amblygonite, lepidolite, petalite, spodumene, alkali tourmaline, triphylite-lithiophilite, zinnwaldite), and Ta (betafite, microlite, tantalite, samarskite) (Berry, 1972; Czerny, 1982). Additional minerals could be listed for these elements, along with a host of other species for As, Bi, C, Ce, Cu, F, La, Mo, Nb, P, Pb, Rb, S, Sb, Sn, Th, U, W, Y, Zn, and Zr, represented either singly or in various combinations.

Textures

The most distinctive features of pegmatites are coarseness of grain and great variations of grain size over short distances (Fig. 2). The average grain size for all occurrences probably is on the order of 10 cm, but mica and quartz crystals 1–3 m across, feldspar crystals 2–10 m across, beryl and tourmaline crystals 2–6 m long, and spodumene crystals 2 m to nearly 15 m long are not uncommon. Individual crystals of allanite, columbite, fluorite, monazite, and

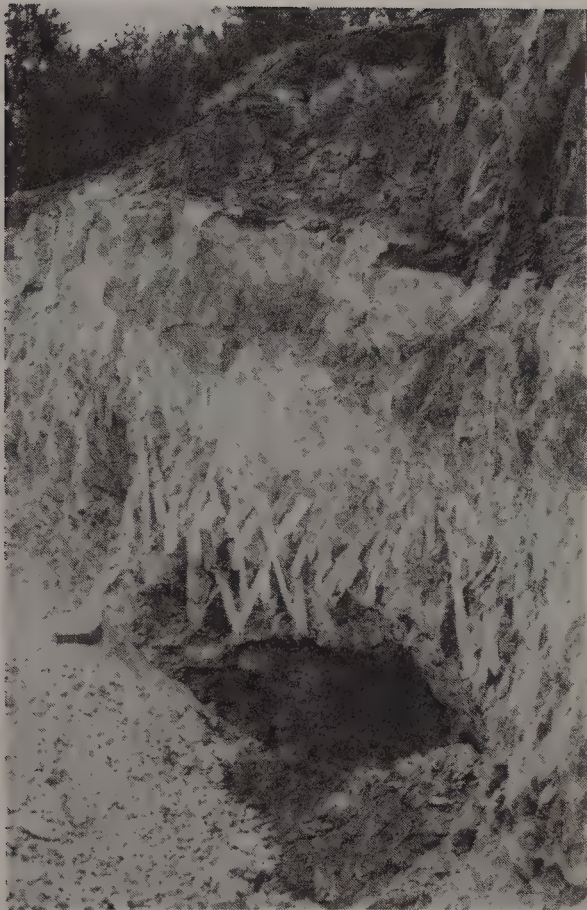


TABLE 1. Grain-size classification for pegmatites based on maximum exposed dimensions of individual crystals

Fine	<2.5 cm
Medium	2.5–10 cm
Coarse	10–30 cm
Very coarse	>30 cm

spectrum of textures. Porphyritic textures are particularly common. Many coarse-grained pegmatites contain interstitial aggregates of relatively fine-grained material, aplitic in appearance, illustrating the affinity of these two rock types. Such complex bodies are often termed “aplopegmatite,” but as this strictly means a “leucocratic or pale-colored pegmatite” (irrespective of texture), its use should be avoided.

The K-feldspar in pegmatites is usually perthitic. The plagioclase ordinarily appears as thin, regularly arranged blades less than a millimeter to several centimeters long. Quartz commonly is intergrown in cuneiform or runic fashion with host crystals of alkali feldspar to form graphic granite. More or less regular intergrowths of hematite, magnetite, or other minerals in muscovite, and of muscovite, garnet, or tourmaline in quartz also are widespread.

Additional complication of texture, found over a broad range of scales, reflects the entry, corrosion, and replacement of some pegmatite minerals by others. This small-scale fracture filling and fracture-controlled replacement is so widespread that it can be regarded as a general characteristic of pegmatites. Replacement is often no more than corrosion along grain boundaries or fracture surfaces, but extensive or even complete replacement of certain minerals by others is by no means rare. Among the best-known examples are masses of coarse albite formed at the expense of quartz and K-feldspar or quartz and muscovite, and pseudomorphs of sugary albite after crystals of K-feldspar or muscovite, scaly lepidolite after spodumene, and fine-grained muscovite after spodumene, topaz, or tourmaline. Also known are some remarkable occurrences of pseudomorphism extending throughout considerable volumes of coarse-grained pegmatite in which replacement has been complete but highly selective, for example, quartz by albite and accompanying spodumene by muscovite.

Graphic or runic textured pegmatite may give the appearance of being relatively fine-grained, whereas the intimately intergrown individual crystals may be structurally continuous over

FIGURE 2. Zonal structure in upper part of a thick, gently dipping sheet of coarse- to very coarse-grained pegmatite beneath dark-colored metamorphic rocks, Harding mine, New Mexico, U.S.A. From top down, the zones consist of (1) quartz with alkali feldspars, apatite, beryl, and muscovite (light band); (2) quartz (darker band); (3) quartz with giant lath-like crystals of spodumene as much as 4 m long (thick band with irregularly comb-like appearance); and (4) quartz, alkali feldspars, spodumene, lepidolite, and microlite (mined-out recess). Below the quarry floor is mainly of finer-grained pegmatite and aplite.

topaz weighing hundreds of kilograms also have been found. The largest crystals ordinarily occur in the interior parts of relatively large pegmatite bodies, but individuals of feldspar, mica, tourmaline, and other minerals extend from wall to wall of some relatively small bodies. However, many large bodies of pegmatite contain no crystals of exceptionally great size. Most pegmatite textures are large scale equivalents of those found in common igneous rocks, and the differences in scale are best treated by means of a special grain-size classification for the pegmatites (Table 1).

Although many pegmatites are equigranular with xenomorphic (anhedral) grains, as with plutonic rocks in general there is a huge

many meters. Such textures are common in complex pegmatites where phase transformation has occurred in the solid state, e.g., the Li-pegmatites of Bernic Lake, Manitoba, Canada, where graphic spodumene + quartz has replaced original petalite (Berry, 1972; Czerny, 1982).

Internal structure

Many pegmatite bodies show considerable internal variations in texture, while remaining compositionally homogeneous, but complex pegmatites (particularly pegmatitic ore deposits) are texturally, compositionally, and mineralogi-

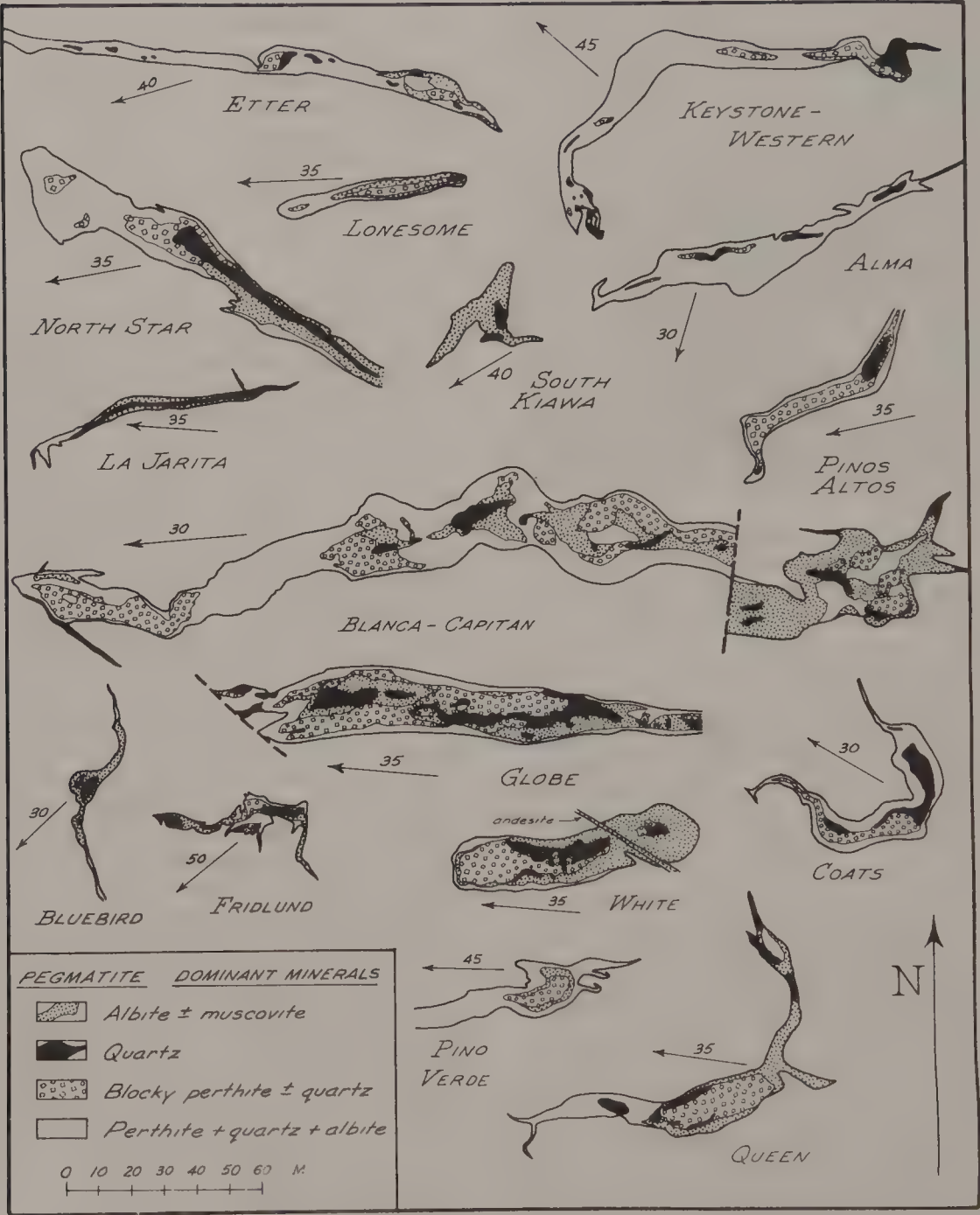


FIGURE 3. Simplified geologic maps of 16 internally zoned pegmatite bodies, Petaca district, New Mexico, U.S.A., showing typical distribution of major lithologic units as exposed at ground surface. Most of the bodies plunge at moderate angles (numbered arrows), in general conformity with prominent linear elements in the enclosing metamorphic rocks, but are otherwise nearly vertical.

cally zoned. Such internal zoning is systematically expressed as successive layers on concentric shells (Figs. 3,4). In horizontal section the zones tend to be arranged about a single or segmented core, but in vertical section they are typically asymmetrical and less continuous. Asymmetry in a vertical sense is especially evident in nearly flat-lying bodies with tabular form, because their zones appear in most exposures as stacks of well-defined layers. For bodies with other forms and attitudes, the deciphering of zonal relationships in three dimensions ordinarily requires exposure by exploration or mining activities.

Fine-grained granitoid pegmatite forms relatively regular and continuous outer zones of many bodies, but it is essentially absent from others. Zones of both finer- and coarser-grained

pegmatite tend to be less continuous and less symmetrically distributed in a vertical sense, regardless of whether they are aplitic, porphyritic, or gigantic in texture. Their distribution is nonetheless systematic, so that most zoned bodies of pegmatite are characterized by relatively coarse-grained, K-rich upper parts, coarse-grained and Si-rich interiors, and relatively fine-grained, Na-rich lower parts. The lower parts of many are aplitic. Associations of major minerals within successive zones consistently follow a definite sequence of worldwide application, and occurrences of varietal and accessory minerals generally can be correlated with certain of these associations. These are features of special significance in the economic appraisal of pegmatites and the working of pegmatite deposits.

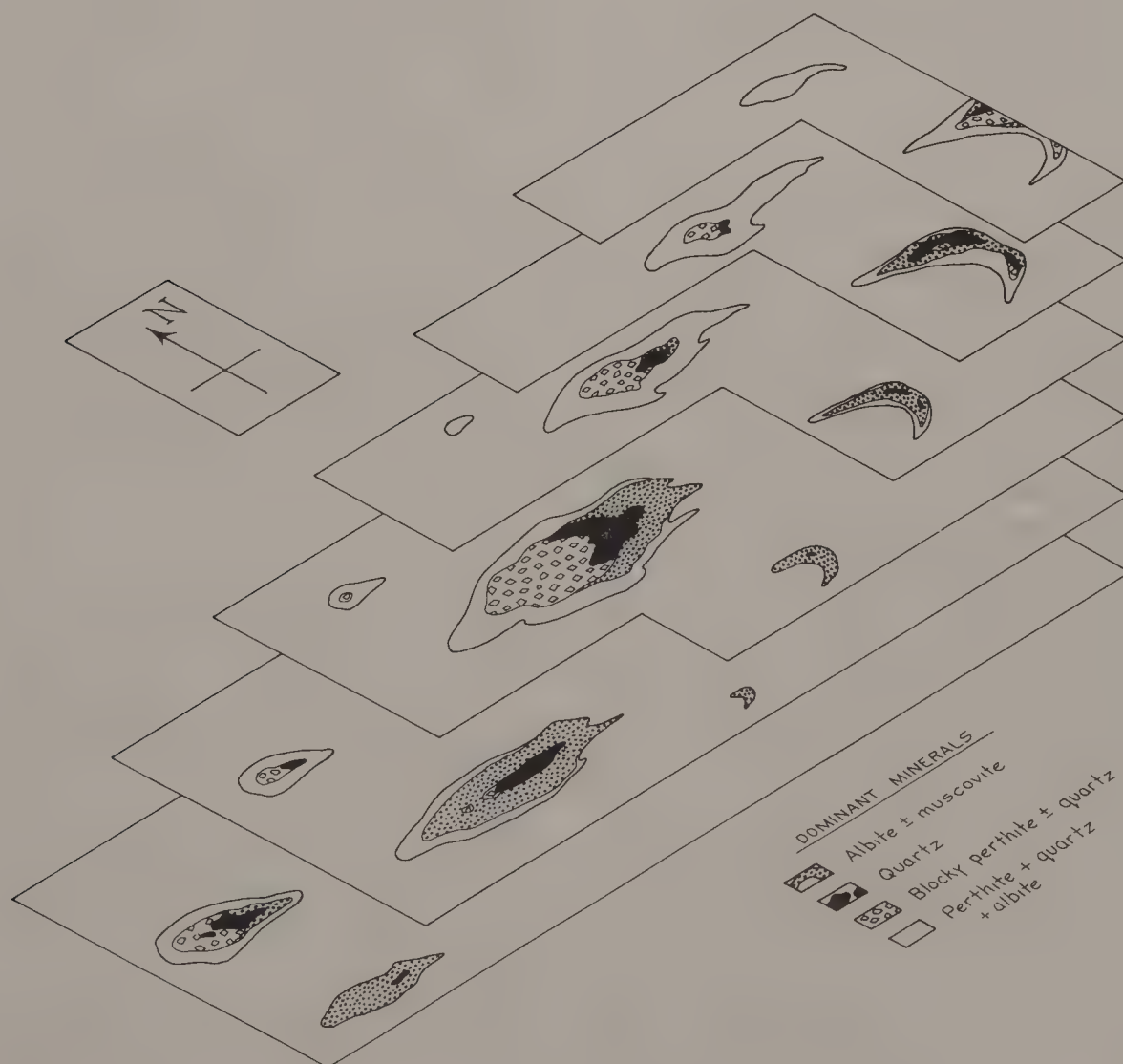


FIGURE 4. Isometric plate diagram of three pegmatite bodies with contrasting shapes and moderately steep westerly plunges, showing general distribution of major internal lithological units (zones); the three-dimensional relationships are typical for the Petaca district of New Mexico, U.S.A. (Fig. 3), and for many pegmatite districts in other parts of the world.

Fracture-filling masses of tabular form, generally a few millimeters to about 3 m thick, transect one or more zones in some pegmatite bodies. They are composed mainly of quartz or of quartz and feldspars, with or without marked concentrations of muscovite or less common minerals. Replacement bodies of lenticular, branching, network, or pod-like form may contain concentrations of muscovite, lepidolite, or other minerals that are found in zones, and notable concentrations of rare minerals may be present as well.

The interior parts of certain pegmatite bodies formed at intermediate or shallow depths in the Earth's crust contain irregular pockets and cavities, generally a few centimeters to several meters in maximum dimension, that are large-scale counterparts of miaroles and miarolitic cavities known in some granitic plutonic rocks. Many of these cavities contain beautifully formed crystals, and such occurrences have yielded transparent gemstones of beryl, garnet, quartz, spodumene, topaz, tourmaline, and other minerals.

Genesis

Pegmatites are quite different from ordinary igneous rocks in many aspects of texture, yet the distribution, structural relationships, bulk composition, rare-element content, and internal features of many pegmatite bodies strongly suggest an origin involving igneous processes. These bodies could well have been formed from the residual fractions of crystallizing melts (magmas) or from the initial fractions of deeply buried crustal materials undergoing partial melting. Such fractions typically would be silicate liquids relatively rich in water, halogens, and other volatile constituents, and also in many of the less common elements. They could remain within the crystallized or partly melted parent rocks to form bodies of segregation pegmatite with irregular and gradational boundaries, or they could be forced away to form homogeneous or zoned bodies of pegmatite elsewhere in the parent rocks or in rocks beyond.

Other pegmatite bodies of evidently metamorphic origin consist of materials probably derived from the enclosing rocks or from more distant sources via aqueous pore fluids, with concentration and crystallization at favorable sites in the form of secretory or replacement masses. Such transfer of materials could well account for numerous veins, pods, and irregular bodies of pegmatite with relatively simple composition and internal structure, and for the development of pegmatite on a grand scale in the Precambrian terranes of Fennoscandia,

India, central Africa, and E Canada.

Studies of pegmatites and experimental investigations of pertinent silicate systems indicate that the unifying factor in pegmatite genesis is the presence of a high-temperature aqueous fluid. When such a fluid is derived by exsolution from a crystallizing water-rich magma, it serves as a low-viscosity medium into which other constituents can be preferentially transferred from the magma, and through which they can diffuse rapidly but selectively in response to temperature-induced concentration gradients. This is an effective vehicle for the nourishment of large crystals and development of coarse textures, and for the gross segregation of constituents to form zonal structure in many pegmatite bodies. Upward gravitational rise of the low-density aqueous fluid can also account for the observed vertical asymmetry (Fig. 4) and especially for preferential development of large crystals in the upper parts of pegmatite bodies while abundant aplite and other finer-grained materials are crystallizing from residual magma in the lower parts.

The aqueous fluid promotes exsolution of alkali feldspars, mobilization of the released albitic constituents, and extensive reactions among earlier-formed crystals both before and after all magma has been used up. Thus, pegmatites of igneous origin reflect metasomatic activity as well. Moreover, the transfer of materials and the mineral replacements implied in the origin of metamorphic pegmatites can be expected to occur if an interstitial aqueous fluid is available in the hot host rocks and a temperature or pressure gradient is present, whether or not a magma is at hand.

Other terms

Aqueo-igneous—refers to the formation of a mineral or rock from magma rich in H₂O, e.g., amygdale, pegmatite.

Bowralite—syenitic pegmatite mainly composed of tabular euhedral sanidine crystals with subordinate arfvedsonite and aegirine (Bowral, N.S.W., Australia).

Laanilite—a coarse-grained pegmatitic rock composed mainly of garnet, biotite, quartz, and iron ores (Laanila, Finnish Lapland).

Micropegmatite—micrographic aggregates of quartz and feldspar occurring as a mesostasis in various igneous rocks.

Pegmatitic stage—a stage in magmatic crystallization during which there is sufficient enrichment in volatiles to permit pegmatite formation.

Pegmatoid—a very coarse-grained facies of an igneous rock having the habit of a pegmatite but lacking the typical granitic composition.

Yatalite—a syenitic or dioritic pegmatite

mainly composed of uraltite, albite, magnetite, and some quartz.

RICHARD H. JAHNS

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* For additional references see **Pegmatitic ore deposits**.

Cross-references: *Aplite; Batholith; Emplacement—modes; Granites and mineralization; Pegmatitic ore deposits; Plutonic rocks—classification and nomenclature; Water in rock melts.*

PEGMATITIC ORE DEPOSITS

Pegmatites as ore bodies are important sources for a number of metals and industrial raw materials, particularly for the ceramic, electronic, and nuclear power industries. Simple

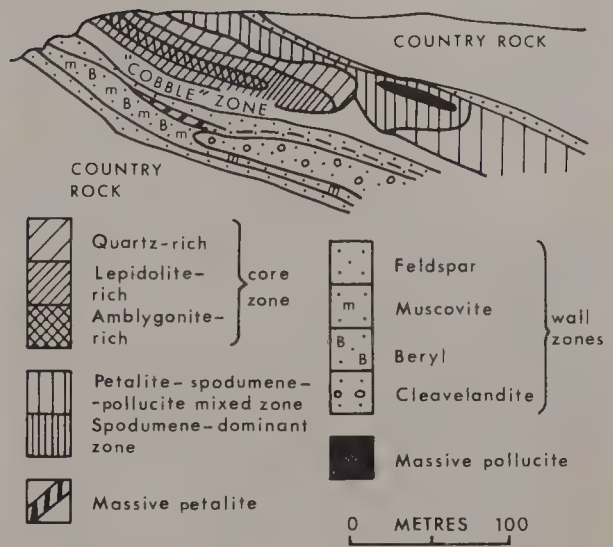


FIGURE 1. Cross-section of zoned, complex Li-bearing pegmatite, Bikita, Zimbabwe. Lepidolite, amblygonite, petalite, and spodumene are Li-minerals; pollucite is a Cs-Na aluminosilicate; cleavelandite is a glassy albite with a tabular habit.

pegmatites, dominated by alkali $\pm$ plagioclase feldspars + quartz $\pm$ biotite $\pm$ muscovite are exploited for their feldspar and mica content. Their overall size and mineral grain size determine whether or not they are exploitable. Complex pegmatites are frequently zoned (Fig. 1), and early crystallized anhydrous silicates often show the effects of later alteration (metasomatism) producing a large and complex suite of boron-, beryllium-, lithium-, and rare metal- and rare-earth element (REE)-bearing minerals. Such pegmatites carry important deposits of Be, Cs, Sn, W, Nb, Ta, Li, U, Th, and REE, e.g., Bernic Lake (Canada), Bikita (Zimbabwe), New Mexico (U.S.A.), Varutrask (Sweden), Greenbushes (Australia), Rajahstan (India), Rubicon (Namibia), and Gravelotte (S Africa). The coarse grain size, and frequently euhedral crystal forms developed also make these deposits sources for both precious and semiprecious gemstones (e.g., emerald, colored quartz, tourmaline, topaz, and spodumene).

The late metasomatic activity in complex pegmatites is marked by the replacement of K-feldspar and plagioclase by albite or white mica, and the appearance of tourmaline and topaz. The fluxing of Li, Be, B, and Cs appears to be achieved by a volatile-rich or alkaline residual melt or fluid in which F^- and PO_4^{3-} are important complexing agents.

The erosion of complex pegmatite-bearing terrane has produced placer concentrations of many minerals, not sufficiently concentrated in the primary pegmatite to be worth exploiting. Such placers are worked for Sn, W, Nb, and Ta in Brazil, and in central and W Africa.

A list of accessory minerals in granitic pegmatites is given in **Pegmatite minerals** (Vol. IVA).

M. ŠTEMPROK

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* For other references and cross-references see **Pegmatite**.

PERALKALINE IGNEOUS ROCKS

In igneous petrology, the term *alkaline* refers to richness in Na and K, but variations in the basis of comparison have resulted in at least five different usages; (1) rocks containing feldspathoid minerals, which are effectively deficient in silica with respect to alkalies (**Alkaline rocks—undersaturated**); (2) rocks containing non-aluminous alkaline minerals, which are effectively deficient in alumina with respect to alkalies; (3) rocks deficient in *both* silica and alumina with respect to alkalies; (4) igneous provinces, magmatic associations, or rock series that are described as “alkaline” according to an arbitrarily defined scale, based on the covariance of $\text{Na}_2\text{O} + \text{K}_2\text{O}$ and CaO, defined by Peacock (1931), or (5) an ill-defined usage stemming from Peacock’s original concept. An historical account of the usages of “alkaline” is given by Sørensen (1974). Usages (4) and (5) include rocks that may fall into categories (1) or (2) or (3). Fortunately, some of the ambiguity has been removed by a recent revival of the description “*peralkaline*” for rocks in category (2). This term, proposed by S. J. Shand, adds precision by accurately specifying an important chemical characteristic that may be found in rocks showing various levels of silica-saturation.

Rocks that are *peralkaline* are distinguished by the presence of non-aluminous alkaline minerals. Typically these are the Na–Fe-silicates

aegirine (acmite), riebeckite–arfvedsonite, or aenigmatite (or solid solutions containing the first two). More rarely, the excess alkalies are expressed in complex Na-silicates such as eudialyte, and even sodium halides. Peralkaline lavas may be wholly or partly glassy; the peralkalinity may then be occult, in which case it is revealed in the chemical analysis as molecular $\text{Na}_2\text{O} + \text{K}_2\text{O} > \text{molecular Al}_2\text{O}_3$, and in the CIPW norm as some combination of acmite, Na-metasilicate, NaCl and NaF. These four constituents are incompatible with normative anorthite, which reflects a natural incompatibility between calcic plagioclase and peralkaline liquids. Peralkaline rocks are characterized by abnormal concentrations of rare elements, such as Zr, and the halogens F and Cl.

Rock types

Peralkalinity is most common in felsic rocks, giving rise to peralkaline varieties of rhyolites (pantellerites and comendites) and granites, trachytes and syenites (see Bailey et al., 1974), and phonolites and nepheline syenites. Some nephelinite lavas, some plutonic rocks of the ijolite group, and malignites are also peralkaline.

Abundance and significance

Compared with basalts and granites, the products of peralkaline magmas are areally and volumetrically minute, but they are important for four reasons. (1) They may contain or be associated with commercially important deposits of Sn, Ta, Nb, rare-earth elements, Th, Zr, Cu, phosphates, and hydrocarbons (Semenov, 1974). (2) They are the only sources of many rare and unusual minerals. (3) They are a characteristic part of the magmatism of the interiors of oceanic and continental plates. (4) The peralkaline condition is opposite to that of the Earth’s crust as a whole, in which alumina is in excess of alkalies. Peralkaline melts must therefore contain some special clues to conditions within the Earth.

Distribution

Peralkaline lavas (rhyolites, trachytes, phonolites) are present in small proportions in most *alkali* basalt provinces, both in the oceanic islands (Borley, 1974, in Sørensen, 1974) and in stable continental areas (Bailey and Schairer, 1966). Some nephelinite volcanoes are also present in the oceans (Bailey, 1974, in Sørensen, 1974), but by far the most spectacular displays of peralkaline magmatism of all types are found associated with uplifted parts of

continental cratons, especially where these are rifted, and it is clear that the uplift, rifting and magmatism are parts of a major process affecting these regions of the lithosphere (Bailey, 1983).

Origin

There are still many unresolved problems connected with the origin of these and other rocks rich in alkalis. Most peralkaline rocks have all the petrographic and petrochemical hallmarks of mantle sources (e.g., stable isotope compositions), but some rhyolites show evidence for a possible origin by melting within the crust. A crucial question is the origin of the peralkaline characteristic. Many petrologists have favored the solution proposed by Bowen (1945), that the crystallization of excess calcic plagioclase leaves a residual melt depleted in alumina—"the plagioclase effect." Such a process could not apply to some peralkaline melts, which must originate in the mantle, at pressures outside the range of plagioclase stability. Even at lower pressures the "plagioclase effect" poses new questions as to why it should operate in some cases and not in others. There is increasing evidence, too, that alkali-rich melts are generated in systems open to volatiles, alkalis, and other mobile elements. It is likely that alkaline and peralkaline melts have been generated by more than one process, but open systems are more compatible with the modern chemical data and the abundant evidence of high volatile emission in alkaline volcanic provinces. A full discussion of peralkalinity, and the possible causes, is given in Bailey (1976).

D. K. BAILEY

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Cross-references: *Alkaline rocks—undersaturated; Basalt; Granite; Igneous rocks—classification and*

nomenclature; Magma; Magmatic crystallization in open systems; Phonolite; Rift-valley volcanism; Syenite; Trachyte.

PERIDOTITE

Peridotite is an ultramafic plutonic rock containing 40–100% olivine (from *peridot*, French for olivine); other essential minerals include clinopyroxene and orthopyroxene; common accessory minerals are hornblende, mica, spinel, garnet, feldspar, ilmenite, and sulfides. Varieties are defined by the presence of phases in addition to olivine; *wehrlite* is olivine + clinopyroxene; *harzburgite* is olivine + orthopyroxene; and *lherzolite* is olivine + orthopyroxene + clinopyroxene. All these varieties contain >40% olivine, and grade to *pyroxenite* which carries <40% olivine. *Dunite* (or *olivinite*—Russian usage) contains >90% olivine. These root names are further qualified by the presence of accessories, e.g., spinel *lherzolite*, garnet *lherzolite*. *Peridotites* carry up to 5% feldspar after which they grade into *gabbro*, *norite*, and *troctolite*. *Cortlandtite* (*hudsonite*) is a *peridotite* containing hornblende and olivine.

Mineral chemistry

Olivine is usually magnesian (Fo_{60-95}); it also carries appreciable Ni (500–5,000 p.p.m.) and Co (50–1,000 p.p.m.). Orthopyroxene is also generally magnesian (En_{60-95}) with variable amounts of Cr, Fe^{3+} , and Ni. Some *peridotites* carry high-Al orthopyroxene. Clinopyroxenes are invariably calcic and both augitic and diopsidic species are common; high-Al species are widespread, while high Ni, Co, and Cr contents are also common. Amphiboles are usually aluminous, pargasitic, or tschermakitic hornblendes with $\text{Mg}:\text{Fe}^{2+}$ ratios > 1. Micas can be either biotite or phlogopite. Spinel shows wide compositional variation, but fall between the end-members chromite–hercynite–spinel–magnetite. The $\text{Mg}:\text{Fe}^{2+}$ ratio is usually > 1, but the Cr:Al ratio varies considerably, possibly defining high-Al and high-Cr groups. Feldspars are invariably calcic plagioclase (An_{60-95}). Garnets are of pyrope–almandine composition, with important amounts of the grossular and uvarovite component. The $\text{Mg}:\text{Fe}^{2+}$ ratios are usually > 1. Sulfides may be locally important, and are usually members of the non-stoichiometric monosulfide solid-solution series, between $\text{Fe}_x\text{S}_{x-1}$ and $\text{Ni}_{x-1}\text{S}_x$ (pyrrhotite–pentlandite) with variable amounts of Co, Zn, and Cu present. Platinide group elements may also be important.

Occurrence

Peridotite is the major component of the Earth's mantle (see **Pyrolite**), and a minor, but important component of the Earth's crust. At the surface of the Earth it occurs in two settings: (1) as primitive material rapidly derived from the mantle, as inclusions in basalts, nephelinites or kimberlites (Carswell, 1980), or as the mantle component of ophiolite complexes; and (2) as an ultramafic cumulate rock in basic and ultrabasic intrusions and in ophiolites.

In the first case the petrology reflects a high-pressure-high-temperature paragenesis in which a series of *P-T* facies can be defined, based on the distribution of Tschermak's molecule between pyroxenes, garnet, and feldspar (O'Hara, 1967). Such rocks include garnet, feldspar, spinel or rarely amphibole peridotite, which represent either pristine mantle (lherzolite) or "depleted" or residual mantle from which a basaltic partial melt has been extracted (harzburgite). These rocks may have either igneous or metamorphic texture (Boullier and Nicolas, 1975).

A related group of rocks are the peridotitic komatiites (Arndt and Nisbet, 1982), erupted as lava, which crystallized as olivine $\pm$ pyroxene $\pm$ spinel + glass. Peridotite has also been recovered from the Moon as blocks and clasts in breccias thought to represent cumulate material from the gabbro-anorthosite lunar highlands.

Other types

Alexoite—a pyrrhotite-bearing dunite (Alexo mine, Dundonald Township, Canada).

Bielenite—contains diallage, enstatite, olivine, chromite, and magnetite with the presence of more pyroxene than olivine distinguishing it from lherzolite.

Eulysite—recorded as a pyroxene peridotite with granular texture, but the presence of Mn-rich fayalite, Mn-garnet, and abundant magnetite, and the association with banded ironstones, point to derivation by metamorphism of the products of hydrothermal leaching and exhalation.

Garéwaite—see lamprophyre.

Gorundite—a garnet-bearing wehrlite composed mainly of olivine and less pyroxene with small proportions of pyrope, picotite, and opaque oxides.

Harrisite—an anorthite-bearing peridotite in which the olivine crystals are oriented approximately at right angles to the cumulate layering, giving *harrisitic texture*.

Kazanskite—a hypabyssal plagioclase-bearing dunite with accessory hornblende and green spinel.

Koswite—olivine-diallage-hornblende perid-

otite with interstitial magnetite (Koswinsky, U.S.S.R.).

Montrealite—a pyroxene-hornblende-olivine peridotite; a highly melanocratic variety of olivine essexite with little or no feldspar or nepheline (Montreal, Canada).

Stubachite—an altered diallage peridotite with tremolite, talc, serpentine, magnetite, pyrite, and breunnerite (Stubachtale, Austria).

Valbellite—a hypabyssal rock composed of bronzite, olivine, hornblende, and magnetite; a microperidotite.

Weigolith—an amphibole- and enstatite-bearing peridotite or microperidotite.

ADRIAN F. PARK

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Cross-references: *Harzburgite*; *Intrusion—layered*; *Lherzolite*; *Oceanic crust*; *Ophiolite*; *Pyrolite*; *Pyroxenite*; *Ultramafic rocks*; *Upper mantle—experimental petrology*.

PETROCHEMICAL CALCULATIONS

Petrochemical calculations aim to manipulate chemical analyses of rocks and sometimes minerals into a form in which the petrologic significance of the data is more apparent than it is in the weight percentages of the oxides in which analytical results are customarily expressed.

Differentiation indices

The simplest procedures give certain ratios, called *differentiation indices*, that have been found, both from experimental syntheses and previous experience with rocks and minerals, to vary systematically with progressive crystallization from magmas, thus enabling the sequence of crystallization of specimens from an igneous body to be determined. The commonest such ratio is the molecular proportion of MgO to molecular proportion of $\text{MgO} + \text{FeO} + \text{Fe}_2\text{O}_3 + \text{MnO}$ in which earlier-formed rocks and minerals have higher values than those formed later. Other ratios commonly used include Rb/K, Rb/Sr, Ba/Rb and the molecular proportion of K_2O to molecular proportion of $\text{K}_2\text{O} + \text{Na}_2\text{O}$. The first step in many petrochemical calculations is to obtain molecular proportions, because these are a measure of relative combining power and the effect of the differing atomic weights of the elements concerned is nullified. Molecular proportions are calculated by dividing the wt.% oxide by the molecular weight of the oxide, which is the sum of the atomic weights of the constituent atoms in the formula unit.

A more complex index is that of Larsen which is $\frac{1}{3}\text{SiO}_2 + \text{K}_2\text{O} - \text{MgO} - \text{CaO} - \text{total Fe}$ (calculated as FeO), all in wt.%, while Nockolds modified this to $(\frac{1}{3}\text{Si} + \text{K}) - (\text{Ca} + \text{Mg})$ in order to deal with iron enrichment separately. Thornton and Tuttle's (1960) differentiation index (or DI) has also been frequently used and this is based on the sum of the calculated quartz + albite + orthoclase + nepheline + leucite + kalsilite, according to the rules of norm calculation.

Norms

A common manipulation recalculates the oxides into terms of minerals using slightly simplified compositions when complex minerals such as pyroxenes are involved. In the most frequently used method, the calculation is into a standard set of commonly occurring minerals, called normative minerals, because they normally occur in igneous rocks (e.g. pyroxene, olivine, feldspar, magnetite, and quartz), irrespective of whether these minerals actually occur in the rocks analysed. The hypothetical assemblage is consequently called a norm or a *CIPW norm* after the originators (Cross, Iddings, Pirsson, and Washington, 1902—see Holmes, 1930, Johannsen, 1931, for the precise method of calculation).

Historically, this method developed as a means of classifying igneous rocks because it was difficult to obtain the exact percentages of each mineral actually occurring in any rock (the

modal composition of the rock) and this encouraged a purely chemical classification. The calculation of a standard set of commonly occurring minerals ("standard minerals") facilitated comparison with other analyses and with actual rocks, bringing the chemical analyses more into the range of petrographers' experience and comprehension and elucidating certain critical petrologic features such as whether feldspathoid, olivine, or quartz would be likely to exist in the rock. A precise measure of the silica saturation of the rock in terms of minerals was thus provided, with oversaturated rocks containing quartz, undersaturated rocks feldspathoid or olivine, and saturated rocks containing none, or very little, of these minerals. Many schemes of igneous rock classification are primarily based on silica saturation, the proportions of K-, Na- and Ca-feldspar and the presence and nature of the pyroxenes (e.g., Shand, 1927—see Johannsen, 1931; Streckeisen, 1976).

In order that similar chemical analyses shall yield similar norms, not only are the standard, or normative minerals alone allowed, but the calculation has to be carried out in a fixed sequence to exclude certain incompatible mineral combinations that are not generally found in igneous rocks. Thus, nepheline and orthopyroxene cannot both occur in any one norm, nor can quartz and olivine or quartz and feldspathoid. Unequivocal distinction between the hypothetical normative minerals and the real or modal minerals was emphasized by the use of standard abbreviations for the normative minerals such as *ab* for albite, *an* for anorthite, and *mt* for magnetite. The actual use of the norm as a means of chemical classification is so complex and the rules are so intricate that such a classification has never been generally adopted, although the primary division of the normative minerals into a silic group (quartz, feldspar, and feldspathoid) and a femic group (pyroxene, olivine, iron ores, and calcite) is widely used in petrology (often rather loosely) with reference to modal minerals and including hornblende and biotite as femic minerals. The use of the norm as a means of distinguishing alkali basalts with normative nepheline from olivine tholeiitic basalts, with normative olivine but lacking nepheline, and tholeiitic basalts without olivine or nepheline, has been firmly established. The method is, moreover, useful in classifying glassy or very fine-grained igneous rocks (such as obsidian), where the norm gives the probable minerals that would have crystallized and also the probable feldspar composition. It is also useful in obtaining the magma type and the probable original mineralogy of metamorphosed basalts now recrystallized into

TABLE 1. Chemical analysis, norms, mode, and Niggli numbers for an amphibolite, Connemara, Eire

Chemical analysis		CIPW norm ^a			Molecular norm ^a			Niggli numbers		
SiO ₂	48.11	or	3.89	59.89	or	4.00	62.88	si	112	
Al ₂ O ₃	16.07	ab	29.87		ab	32.45		al	22	
TiO ₂	1.74	an	26.13		an	26.43		fm	48.5	
Fe ₂ O ₃	2.90	wo	6.03	11.68	wo	5.84	11.68	c	21	
FeO	7.97	en	3.80		en	3.68		alk	9	
MgO	7.97	fs	1.85		fs	2.16		ti	3.03	
CaO	8.35	en	2.51	3.43	en	2.20	3.50	p	0.19	
Na ₂ O	3.55	fs	0.92		fs	1.30		k	0.11	
K ₂ O	0.67	fo	9.94	15.24	fo	10.09	16.02	mg	0.57	
MnO	0.16	fa	5.30		fa	5.93				
P ₂ O ₅	0.19	ilm	3.34		ilm	2.46				
H ₂ O +	2.66	mt	4.18		mt	3.09				
	100.34	ap	0.34		ap	0.37				
Normative feldspar		or	ab	an	or	ab	an			
		6.5	49.9	43.6	6.4	51.6	42.0			
Modal analysis ^b										
Hornblende	55.0	Chlorite	6.0		Sphene	tr.				
Plagioclase (An ₄₀)	35.5	Iron ore	2.5		Epidote	2.0				
Quartz	tr.	Apatite	tr.							

^a Molecular norm (kata-norm) calculated to the same normative minerals as CIPW norm; both contain olivine but no nepheline and the original basaltic type was therefore an olivine tholeiite whose original minerals were probably similar to the normative ones calculated; normative plagioclase is notably albite-rich for a basalt.

^b Modal analysis shows the mineralogical composition of the rock together with the optically determined composition of the plagioclase, which is close to the normative feldspar composition.

hornblende-plagioclase rocks (e.g., amphibolites). An example of this application is given in Table 1.

A development from the CIPW norm was the *molecular norm* suggested in 1936 by Niggli (see Niggli, 1954). The method offers a choice between calculating into a standard set of minerals, which for igneous rocks is the same as the CIPW standard set, or calculating the analysis into terms of the modal minerals or any other chosen minerals, including biotite and hornblende, which, although of very common occurrence, were omitted from the CIPW system because of their relatively complex formulae. The minerals are calculated in molecular percentages instead of the previous wt.% because molecular mineral percentages are much closer to volume percentages of minerals than are weight percentages and, in addition, the molecular minerals are more readily transformed into alternative minerals. This more flexible approach allows application to metamorphic rocks for which classification has always been simple and based on modal minerals. Here the prime need is to calculate what mineral assemblage any given rock—sedimentary, igneous, or metamorphic—can give if metamorphosed to any particular grade

of metamorphism or what any metamorphic rock could have been before metamorphism. Thus, given a shale or slate analysis, it is possible to calculate how much sillimanite, garnet, quartz, biotite, and feldspar it will recrystallize into under, for example, the highest grades of metamorphism (the katazone). Such a norm is termed a kata-norm if calculated in accordance with set rules, or a kata-variant if particular high-temperature minerals are chosen. The procedure for medium grades of metamorphism gives a mesonorm while the epi-norm is the composition stated in terms of low-grade metamorphic minerals such as chlorite and sericite. A detailed exposition of the method is given by Burri (1964).

The advantage of choosing normative minerals (as the calculated minerals are still called), that are the same as the modal minerals is that it is possible to calculate the composition of the modal minerals in many instances, if the chemical and modal compositions of the rock are known and the composition of some of the minerals is fixed (e.g., quartz) or known from optical properties or chemical analysis (e.g., plagioclase, olivine, orthopyroxene). In this way, mineral compositions may be obtained without separating and analysing the minerals

themselves, although the accuracy of the result is unacceptably low unless the mineral is abundant. The availability of electron microprobes means the method is most useful for minerals with complex exsolution or zoning, where the bulk composition is not readily obtained by microprobe analysis. Many authors choose the modal minerals to be normative minerals with the mistaken idea of demonstrating the accuracy of the chemical analysis by obtaining close agreement between the norm and the mode. This is only acceptable if the

compositions of the modal minerals are known and these same compositions are employed in calculating the normative minerals.

Variation diagrams

A common petrochemical procedure is to manipulate analyses into a form in which they can be plotted so that the significant variations and relationships of numerous chemical analyses can rapidly be appreciated visually. Quite apart from the hundreds of thousands of already

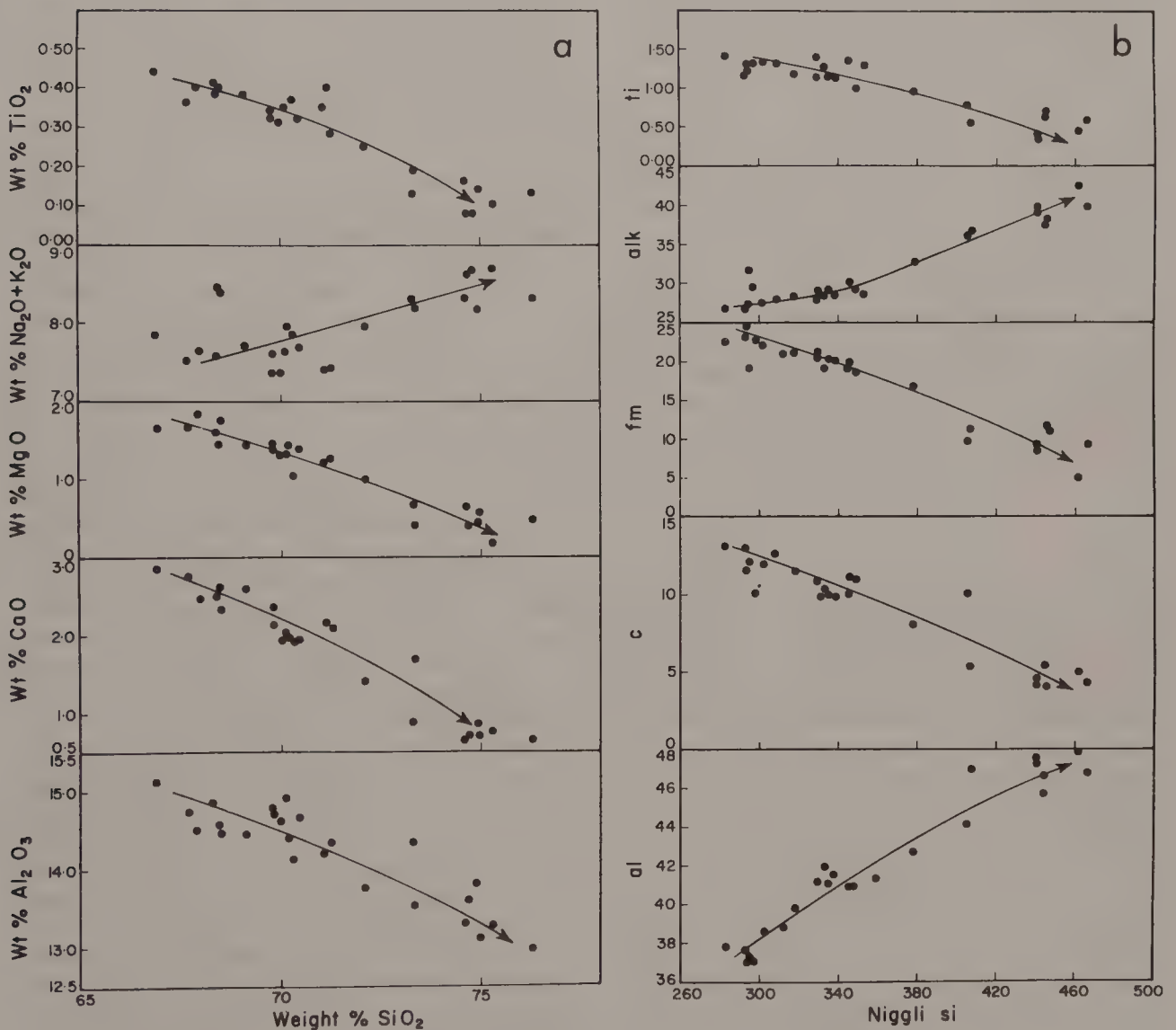


FIGURE 1. Variation diagrams for part of the Galway granite, Eire. Each point is one rock analysis; arrowed lines show differentiation trends, the early-crystallized rocks plotting at the left-hand side of each diagram with the lowest SiO₂ contents while the last differentiates plot near the arrowhead; these latter compositions crystallized at lower temperatures than those with less SiO₂; the fact that the points in both diagrams are roughly arranged in lines suggests that crystallization took place from one parental magma. (Data from Wright, 1964.) (a) Harker diagram; all constituents decrease with increase of SiO₂ except for Na₂O + K₂O. (b) Niggli plots of the same data shown in (a). *al* increases with increase of *si*, owing to increase in the proportion of feldspar at the expense of femic minerals (hornblende, biotite, etc.)—this is entirely obscured in (a) by the constant summation effect; the feldspar becomes less anorthitic and therefore *c* declines, while *alk* in albite and K-feldspar increases. In both (a) and (b) Ti falls because this is chiefly located in hornblende, biotite and sphene.

published rock analyses, with the introduction of instrumental and automatic methods of analysis, the significance of the results from even a single project is not usually apparent without some form of graphical or computer processing. Numerous methods have been proposed but only a small number are commonly used, and these vary from simple plotting of wt.% of an oxide against SiO_2 wt.% (Harker diagrams—see Fig. 1), or MgO (Cox et al., 1979), or against the Larsen or Nockolds index (Fig. 2), to complex plots involving two adjoined projections of a vector whose orientation, length, and position on the two projections indicate different petrochemical properties including certain normative mineral proportions

(Zavaritsky diagrams). Differentiation indices are most often employed as one of the axes of the graph.

The most widely used European method is that of Niggli (1954), first suggested over 50 years ago and developed from a more complex earlier version by A. Osann that it superseded (Johannsen, 1931). *Niggli numbers* are obtained by making the molecular proportions of Al_2O_3 (designated *al*) + Fe_2O_3 + FeO + MgO + $\text{MnO}(\text{fm})$ + $\text{CaO}(\text{c})$ + Na_2O + $\text{K}_2\text{O}(\text{alk}) = 100$. Thus, *al*, *fm*, *c*, and *alk* numbers are molecular proportions given as a percentage of the total molecular proportion (*T*) of these oxides, independent of the remaining constituents of the rock, especially SiO_2 , an important improvement over the methods of Osann and Zavaritsky. The SiO_2 number (*si*) is the silica molecular proportion expressed as a percentage of the same total *T*, while the ratio of the different alkalis (*k*), is molecular $\text{K}_2\text{O}/(\text{Na}_2\text{O} + \text{K}_2\text{O})$ and the Mg/Fe ratio (*mg*) is molecular $\text{MgO}/(\text{MgO} + 2\text{Fe}_2\text{O}_3 + \text{FeO} + \text{MnO})$ —see Table 1.

Like most other systems, the Niggli method groups elements that commonly substitute for each other in silicates but it avoids combining the Al_2O_3 with Na_2O , K_2O , and CaO in the proportions found in feldspar, which is a combination of great use only in igneous petrology, and can cause confusion when applied to many other rocks, although both Osann and Zavaritsky use such a procedure. The exclusion of SiO_2 from the molecular proportions that are summed avoids the inherent negative correlation that exists between most oxides and SiO_2 because increases in SiO_2 —the largest constituent in most analyses—inevitably lead to decreases in other oxides, since the constant sum of 100% for the chemical analysis must be retained. For this reason, Harker diagrams have been strongly criticized (e.g., Chayes, 1971). Many systems have tried to include some features of the CIPW norm, whereas Niggli's system enables these to be readily obtained from Niggli numbers, if desired, but does not have these features built into the numbers themselves. Thus if $al > alk$, as is usual, then silica saturation = $si - 100 + 4alk$ and if this is positive and the rock is not peraluminous ($al > alk + c$) or peralkalic ($alk > al$) then total feldspar = $al + alk$, alkali feldspar = $2alk$, albite/anorthite = $(1 - k)2alk/al - alk$. Niggli numbers have therefore been widely used for petrochemical studies of sedimentary, metamorphic, and igneous rocks (Fig. 1).

Zavaritsky's system is widely used in the U.S.S.R. and the precise method of calculation is given by Zavaritsky and Sobolev (1964, p.

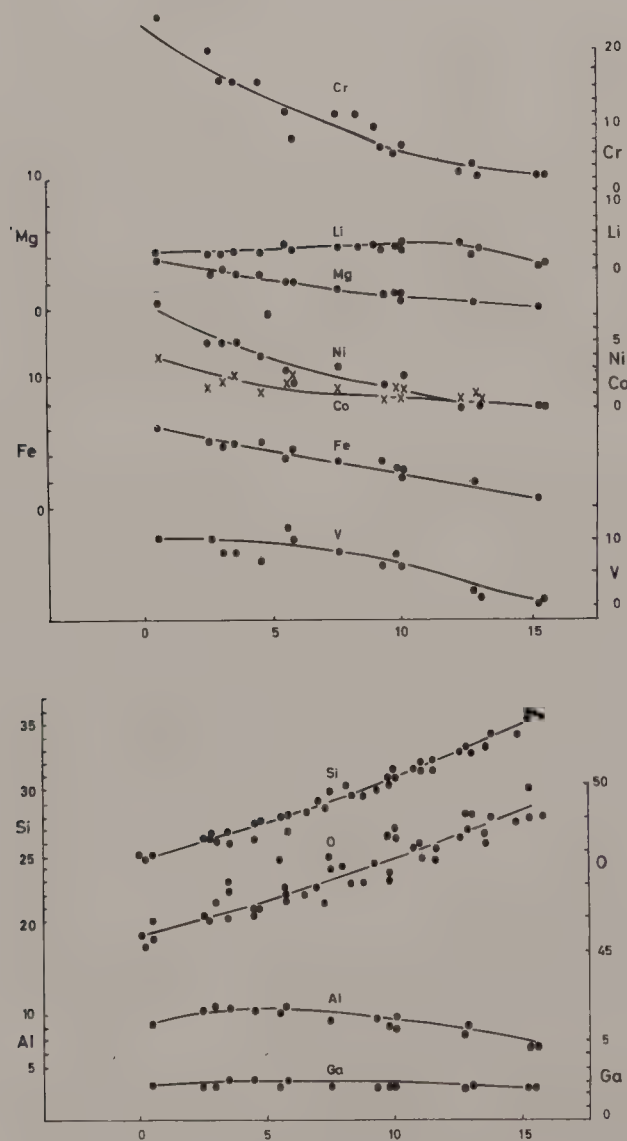


FIGURE 2. Variation diagrams for the igneous rock series of the Scottish Caledonian granites in wt.% (major elements) and parts per million (trace elements) plotted against $(\frac{1}{3}\text{Si} + \text{K}) - (\text{Ca} + \text{Mg})$, the Nockolds differentiation index. (Data from Nockolds and Allen, 1953.)

412–414). There are four principal molecular ratios, a , c , b , and s and by definition $a + c + b + s = 100$. For most rocks, a is the molecular proportion of $\text{Na}_2\text{O} + \text{K}_2\text{O}$ combined with Al_2O_3 , c is that of CaO bound to Al_2O_3 , b is the remaining metal molecular proportions excluding SiO_2 and TiO_2 that form s . For normal rocks $a:c:b:s$ is equivalent to $2alk:(al - alk):(fm + c - (al - alk)):si$ in terms of Niggli numbers. Secondary characteristics include many of the more important features of most norm calculations.

Triangular graphs with their requirement of addition of the three components to 100 have been used by many petrologists. Examples include normative albite, orthoclase, anorthite; Ca, Na, K wt.% (Fig. 3); AFM((Na + K):

$(\text{Fe}^{2+} + \text{Fe}^{3+}):\text{Mg}$ —Fig. 3), triangular slices through the al , c , alk , fm tetrahedron or the di, olv, hy normative plot. Details of how to plot on triangular diagrams are given in Holmes (1930) and Niggli (1954). The system of von Wolff (1922—see Johannsen, 1931) is also based on the use of a triangle and calculates each analysis into an ideal feldspar–augite mixture to form the L and M components while the excess or deficit of SiO_2 , positive or negative, forms the third component Q .

Plots of numerical characteristics of rock series enabled the more precise definition of magma provinces, magma types, and magma series such as the calc-alkali series (gabbro, diorite, tonalite, granodiorite or basalt, andesite, dacite, rhyodacite) while sedimentary, metasomatic, and igneous trends of variation reflecting the different processes involved in forming these rocks became apparent. The search for and recognition of parental or primary magmas, differentiation trends, liquid lines of descent (or lineages), and accumulative rocks (Fig. 3) became much more systematic and quantitative modeling developed by graphical (Wilcox, 1979) or computer (Miesch, 1976) methods. The development of purely numerical concepts took place, such as Peacock's alkali–lime index, which is the SiO_2 wt.% at which the crossing of the CaO and the $\text{Na}_2\text{O} + \text{K}_2\text{O}$ wt.% occurs on a variation diagram.

Mineral calculations

Calculation of the mineral into percentages of end-members is used for feldspar, which has three common end-members showing partial or complete solid solutions between each other: albite ($\text{NaAlSi}_3\text{O}_8$), K-feldspar (KAlSi_3O_8 —often the polymorph orthoclase), and anorthite ($\text{CaAl}_2\text{Si}_2\text{O}_8$). Compositions are expressed cryptically as $\text{Ab}_{70}\text{Or}_{25}\text{An}_{25}$, in which the numbers sum to 100 (see Table 1). Likewise, olivine $(\text{Mg}, \text{Fe})_2\text{SiO}_4$ and orthopyroxene $(\text{Mg}, \text{Fe})\text{SiO}_3$ each have two principal end-members with a solid-solution series between them (forsterite–fayalite and enstatite–ferrosilite, respectively) and their compositions are commonly expressed as $\text{Fo}_x\text{Fa}_{100-x}$ and $\text{En}_x\text{Fs}_{100-x}$. Usually the calculation is in terms of molecular % of the end-members, derived using the molecular proportions, but wt.% is common if the information is to be compared or plotted on a phase diagram constructed by laboratory synthesis. Atomic percentage of the critical elements (e.g., Fe and Mg in olivines and orthopyroxenes or Ca, Fe, and Mg in clinopyroxenes) is also sometimes used. Garnets and iron ores are two other mineral groups with extensive solid-solution series that are

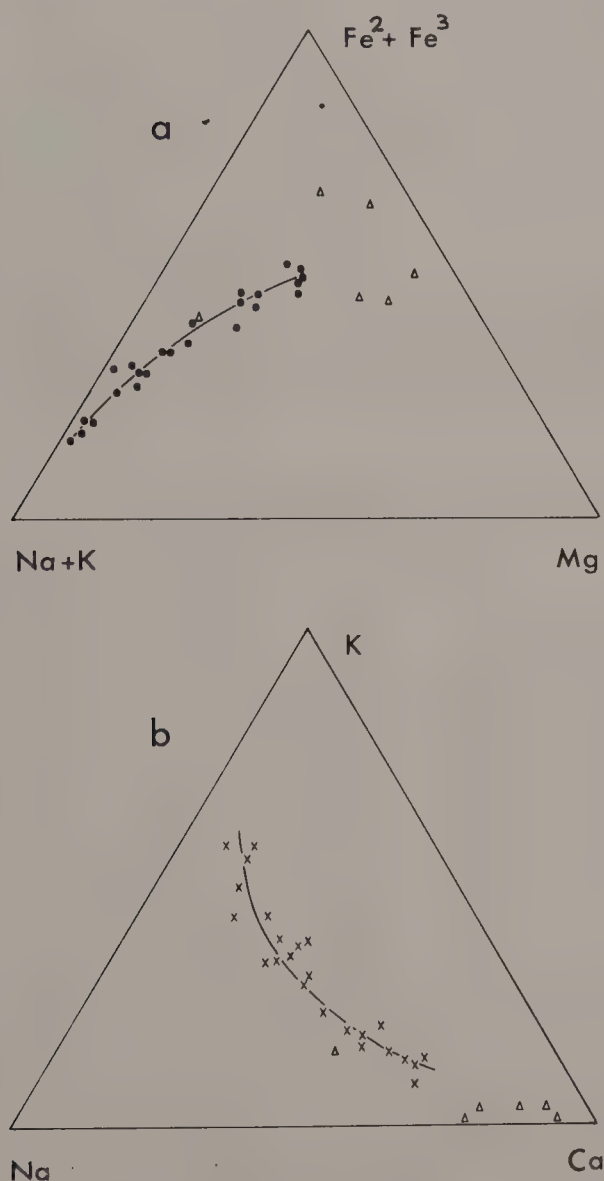


FIGURE 3. Plots of (a) $(\text{Na} + \text{K})-(\text{Fe}^{2+} + \text{Fe}^{3+})-\text{Mg}$ (AFM diagram) and (b) $\text{Ca}-\text{Na}-\text{K}$; Δ , accumulative rocks that do not plot on general trend lines that are supposed to show lines of liquid compositions. (Data from Nockolds and Allen, 1953.)

commonly calculated into proportions of end-members. Clinopyroxenes present special difficulties and both the choice of end-members and the calculation sequence are critical: different methods can give slightly or considerably different results. These difficulties become so acute with the amphiboles, because of the large number of complex end-members that can be chosen, that the method has never been successfully applied in this mineral group.

In general, however, the method enables a much simpler expression of the composition than with the full chemical analysis in oxides, significant variations are easier to appraise, the composition can often be conveniently plotted and nearly all diagrams relating optical properties to composition employ end-members. Consequently practically all petrologists think in terms of end-member compositions for most of the common anhydrous silicates and not in terms of weight percentages of oxides.

Calculation of the number of cations combined with a fixed number of anions, usually oxygen or O + OH + F + Cl is commonly carried out. The fixed number is obtained from X-ray determination of the number of formula units in the unit cell of the mineral: sometimes a simple multiple or fraction of this number is used for convenience. Thus, the unit cell of many amphiboles contains 48 (O + OH + F + Cl) and so the cations are computed on the basis of this

number of anions, or more often of 24 anions, the half-unit cell, on which basis the cations add up to a value between 15 and 16 in most amphiboles (see Table 2 and Deer et al., 1966). In practice, the calculation is often on the basis of the formula unit; thus, olivines ((Mg, Fe)₂SiO₄) are calculated to 4(O) or 8(O) while pyroxenes ((Ca, Mg, Fe)₂Si₂O₆) are often calculated to 6(O).

The method enables the allocation of the different cations to the tetrahedral (usually Si and Al), octahedral (Al, Fe, Mg, Mn, Ti), or larger sites (Ca, Na, K)—often abbreviated to Z, Y, and X sites—as established by X-ray work and thus provides a partial check on the analyses in that sites are not usually grossly over- or underfilled in good analyses. Coupled atomic substitutions are more easily identified and if the cell size is known the density can be calculated very precisely using the Avogadro number, and checked against the determined density; if the latter cannot be determined, the calculated density may be the only available value.

Rarely, the calculation is to a fixed number of cations, which is usually, but not necessarily, an integer. This procedure is particularly useful in the computation of mineral compositions from rock and modal analyses where the allocation of the OH is usually quite uncertain. Thus, 15.5 cations is very close to the contents of the half

TABLE 2. Amphibole composition; mineral from rock in Table 1^a

Chemical analysis		Ions to 24(O, OH) (half unit cell)		Based on 15.5 cations			
SiO ₂	45.94	Si	6.68	8.00	Si	7.2	8.00
Al ₂ O ₃	10.73	Al ⁴	1.32		Al ⁴	0.8	
TiO ₂	1.43	Al ⁶	0.52	5.13	Al ⁶	0.7	5.1
Fe ₂ O ₃	2.85	Ti	0.16		Ti	0.2	
FeO	11.35	Fe ³	0.31		Fe ³	0.2	
MgO	12.65	Fe ²	1.38		Fe ²	1.2	
CaO	10.61	Mn	0.03		Mn		
Na ₂ O	1.38	Mg	2.73		Mg	2.8	
K ₂ O	0.58	Ca	1.65	2.14	Ca	1.5	2.4
MnO	0.22	Na	0.39		Na	0.7	
P ₂ O ₅	0.02	K	0.10		K	0.2	
H ₂ O +	2.23	OH	2.15		OH	1.4	
	99.99		21.85			22.6	
		Σz	8.00	15.27	Σz	8.0	15.5
		Σy	5.13		Σy	5.1	
		Σx	2.14		Σx	2.4	
		mg	0.61		mg	0.66	

^aFrom the modal analysis, the rock analysis and the optically determined plagioclase composition, and assuming the iron ore is two-thirds magnetite, one-third ilmenite, the calculated composition of the hornblende, based on 15.5 cations, is given. From the sum of the cation charges and knowing that O + OH = 24, the OH and O can be calculated. Comparison with the previous column shows that the composition agrees reasonably well for some elements but poorly for others.

unit cell of most amphiboles, while olivines are close to 3 and pyroxenes to 4.

Calculation of impurities out of analysed minerals may be feasible if the formula of the mineral is rather inflexible or the impurities contain elements that do not enter the structure of the mineral analysed. Thus, substantial feldspar impurity in olivine can be corrected for as neither Na nor K enter olivine while Ca and Al are generally very low. Quartz inclusions in garnets are a common source of contamination, but as garnets approximately follow the formula $3R^{4+}O_2 \cdot R_2^{3+}O_3 \cdot 3R^{2+}O(3SiO_2 \cdot (Al, Fe, Mn, Ca)O)$ the excess SiO_2 is readily identified using either of the other two groups as a basis. Apatite with its high P content, is another mineral easily calculated out of most silicate minerals.

Other procedures

The *standard cell* of Barth (1948—see Barth, 1952) is based on oxygen forming ca. 92% by volume of most rocks and the view held at one time by many petrologists that metasomatic replacements took place without a change in volume (the oxygen content of metasomatized rocks was considered as essentially constant, the variation taking place in the remaining 8% of the rock by movement of the cations). Accordingly, Barth proposed that in computing

metasomatic changes a standard cell of 160 oxygens be used as the reference basis on which to calculate the cation changes: 160 oxygens were chosen because most rocks have close to 100 cations combined with 160 oxygen ions and so this facilitated comparison of the cation abundances with the more familiar oxides and molecular percentages already used by petrologists. At the same time it provided cation contents close enough to percentages to be loosely regarded, for inspection purposes, as percentages (Table 3). However, now that oxygen isotope studies have demonstrated the appreciable mobility of oxygen and the massive movement of water and carbon dioxide through metamorphic rocks is well established, the theoretical basis for supposing that the total oxygen in rock remains constant in metamorphism must be regarded as unsound.

The *oxidation ratio* (molecular proportion of $2Fe_2O_3 \times 100 / (2Fe_2O_3 + FeO)$) is a useful ratio which measures the degree of oxidation of iron.

ACF diagrams have been widely used in the study of mineral assemblages in metamorphic rocks in relation to differing conditions of metamorphism. They form a basis for comparing the mineral assemblages actually present with those predicted for the determined rock composition under the supposed conditions of metamorphism. ACF diagrams are essentially

TABLE 3. Standard cell of a hornblende minette (lamprophyre)

	A	B	C	D	E	F
SiO ₂	52.68	0.877	2	1.754	98.85	49.4
TiO ₂	2.71	0.034	2	0.068	3.83	1.9
Al ₂ O ₃	14.02	0.275	1.5	0.412	23.23	15.5
Fe ₂ O ₃	5.04	0.063	1.5	0.094	5.38	3.5
FeO	6.40	0.089	1	0.089	5.02	5.0
MgO	3.35	0.083	1	0.083	4.68	4.7
CaO	6.42	0.114	1	0.114	6.42	6.4
Na ₂ O	2.43	0.078	0.5	0.029	2.20	4.4
K ₂ O	3.30	0.070	0.5	0.035	1.97	3.9
P ₂ O ₅	1.30	0.018	2.5	0.045	2.54	1.0
H ₂ O	1.89	0.210	0.5	0.105	5.92	11.8
	99.54			2.839	159.99	107.5

A Weight % oxides.

B Atomic proportions of each oxide, calculated by dividing SiO₂ by 60.06, TiO₂ by 79.90, Al₂O₃ by 50.97, Fe₂O₃ by 79.85, FeO by 71.85, MgO by 40.32, CaO by 56.08, Na₂O by 30.99, K₂O by 47.10, P₂O₅ by 70.97, and H₂O by 9.01, these values being the molecular weights of the oxides divided by the number of cations in the oxide formula, e.g., 1 for SiO₂, 2 for Al₂O₃.

C Number of oxygen atoms combined with one cation in the oxide formula.

D Column B multiplied by column C.

E Number of oxygen atoms when the sum of column D is recalculated to sum to 160, i.e., column D multiplied by $160 / (2.839) = 56.358$.

F Standard cell giving the number of cations associated with 160 oxygen atoms, obtained by dividing column E by the number of cations associated with each oxygen, i.e., 1/2 for SiO₂, 2/3 for Al₂O₃, 2/5 for P₂O₅, 2 for H₂O.

triangular plots (i.e., $A + C + F = 100$) based on wt.% of $A = \text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3 - (\text{Na}_2\text{O} + \text{K}_2\text{O})$; $C = \text{CaO}$, and $F = \text{MgO} + \text{FeO} + \text{MnO}$ after certain corrections have been applied (see **Metamorphic rock composition—graphical representation**).

*Comparisons of the element patterns of rocks or minerals with the pattern of a reference material is quite common in dealing with the rare-earth elements (REE) or the “incompatible” elements—i.e., those (such as K, Ba, Rb, Ti, Nb, Ta, Th) that remain in the magma preferentially because they do not enter the solid crystallizing phases as easily as the compatible elements (such as Si, Al, Ca, Fe, Mg, Cr, Ni). The resulting plot (“spidergram” or spider diagram) is based on dividing the concentration of each element by the concentration of that element in the reference material, the resulting values being termed “normalized values.” With the REE (Fig. 4) the reference material is the average of a number of chondrite meteorite analyses. In the case of the incompatible elements, a number of reference materials have been used, such as typical mid-ocean ridge basalt (MORB) (see **Flood basalt**).*

The CMAS quaternary system is often used particularly in dealing with ultrabasic and basic rocks: the details of the calculations have been summarized by O'Hara (1976). The scheme

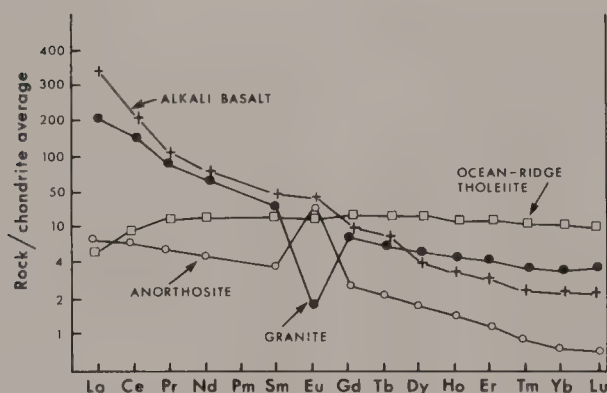


FIGURE 4. A typical representation of the REE data for four rocks normalized to average chondrite meteorite; note the Eu anomaly which is usually positive for plagioclase-rich rocks and negative for rocks derived from magmas that have previously undergone substantial plagioclase crystallization, depleting the magma in Eu. Alkali basalts show enrichment of light REE relative to heavy REE, possibly because of the stability of garnet in the parent mantle rock that partially melted to give the basalt. Typical mid-ocean ridge basalt (ocean-ridge tholeiite) is depleted in the lighter rare earths and has a generally flat middle to heavy rare earth concentration. Generally, in differentiation from basic to acid rocks the concentrations increase, so that the lines move upwards.

converts rock and mineral analyses to four composite components $\text{XO}-\text{YO}-\text{R}_2\text{O}_3-\text{ZO}_2$, thereby reducing the complexity of the chemical data so that it can be represented in a single quaternary system. The components are calculated in terms of equivalent weights of CaO , MgO , Al_2O_3 , and SiO_2 (hence CMAS) and the system is therefore particularly suited for representing four-phase ultrabasic rocks with olivine, two pyroxenes, and an aluminous phase (anorthite or spinel or garnet). The calculation would be much less cumbersome and flexible if it were entirely in moles but because most experimental petrological phase diagrams have been reported in wt.%, the calculation is first done in moles and then finally calculated into the weight of an equivalent number of moles of the component concerned. This enables ready comparison with cotectics, eutectics, and phase boundaries. Thus, if these are 0.45 mole of components such as $\text{FeO} + \text{MnO}$ considered to behave like MgO in silicate substitution, then they are converted into equivalent wt.% of MgO (i.e., $0.45 \times \text{molecular weight of MgO}$) in the final coordinates. Hence, all geochemically similar minerals plot in the same place in the $\text{CaO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2$ system. Corrections have to be made for ilmenite and apatite and to allow Na and K to be calculated in equivalent amounts of anorthite.

The complete calculation of the four components is

XO

$$= 100 (\text{mol of } \text{CaO} - 3.5 \text{P}_2\text{O}_5 + 2\text{Na}_2\text{O} + 2\text{K}_2\text{O}) \times 56.08/\text{SUM}$$

YO

$$= 100 (\text{mol of } \text{MgO} + \text{FeO} + \text{MnO} + \text{NiO} - \text{TiO}_2) \times 40.32/\text{SUM}$$

R_2O_3

$$= 100 (\text{mol of } \text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3 + \text{Cr}_2\text{O}_3 + \text{TiO}_2 + \text{Na}_2\text{O} + \text{K}_2\text{O}) \times 101.96/\text{SUM}$$

ZO_2

$$= 100 (\text{mol of } \text{SiO}_2 - 2\text{Na}_2\text{O} - 2\text{K}_2\text{O}) \times 60.08/\text{SUM}$$

where 56.08, 40.32, 101.96, and 60.08 are the molecular weights of CaO , MgO , Al_2O_3 , and SiO_2 , respectively, and the SUM is the total of all four components.

To plot this four-component system on triangular diagrams requires that the diagrams be compositional planes through the tetrahedron, i.e., projections, so that the variation of one component is considered to be insignificant

for the particular variations studied (see **Geochemical modeling in petrogenesis**).

Other terms

Agpaitic coefficient—the ratio $na + k/al$, where na , k and al are the relative proportions of Na, K, and Al atoms, respectively, in a rock; used especially for alkaline igneous rocks.

Crystallization index (C.I.)—the number that is calculated from the anorthite–diopside–forsterite system and represents the sum (in wt.%) of normative anorthite, Mg-diopside, normative forsterite, normative enstatite converted to forsterite, and Mg-spinel calculated from normative corundum in ultramafic rocks; used in igneous petrology (Poldervaart and Parker, 1964).

Oxygen ratio—the ratio of the number of atoms of oxygen in basic oxides in a mineral or rock to the number of atoms of oxygen in SiO_2 .

Quartz index—the qz Niggli value (number); it may be positive or negative.

Saturation line—the line representing saturation with respect to SiO_2 on a variation diagram for an igneous rock series.

Silica coefficient—the ratio of total silica in a rock to the silica in the feldspars and metasilicates; used in Ossnan's chemical classification of igneous rocks.

BERNARD E. LEAKE

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Cross-references: *Differentiation index*; *Flood basalt*; *Geochemical modeling in petrogenesis*; *Igneous rocks—classification and nomenclature*; *Magmatic differentiation*; *Metamorphic rock composition—graphical representation*; *Statistics in petrology*.

PETROGRAPHIC METHODS

Physical and mineralogical methods used in the study of rocks and minerals constitute *petrographic methods*. They include separation of the mineral constituents of rocks, determination of specific gravity, proportions of various minerals present in the rock (mode), optical properties of minerals using the polarizing microscope, and the preferred orientation of the mineral grains in the rock fabric (Holmes, 1930; Hutchison, 1974; Zussman, 1977).

Mineral separation

Rocks have to be crushed to the appropriate size to free the mineral particles, which are then separated by one of several methods. Separations by gravity use heavy liquids such as bromoform (specific gravity 2.89), methylene iodide (3.25) and Clerici solution (4.28). Mechanical panners and shaking tables, froth flotation, and elutriation methods are also used. Magnetic and electrostatic methods are used when the mineral being separated is susceptible to those properties. In several of the methods, special techniques are used to avoid fine material sticking and contaminating the mineral grains being separated. Hand picking of individual mineral grains, under a binocular microscope, is by far the most tedious of all the methods and is used as a last resort. A detailed

account of mineral separation techniques is given in Zussman (1977).

Specific gravity

Specific gravity is defined as mass per unit volume compared to that of water and determination of it invariably utilizes the Archimedes principle. The weight of the sample (rock or mineral) in air divided by the loss of weight in water gives specific gravity. Several different types of balances—Walker's steel yard, Jolly spring balance, and varieties of analytical balances—are used. Porous rocks need to be crushed to obtain the specific gravity of their constituent rock material. The pycnometer method used in determining the specific gravity of powdered material is often employed when pure minerals are available in powder form from mineral-separation techniques.

Mode

The mode is the proportion (volume % or wt.%) of the different kinds of minerals present in a rock. The estimation (*modal analysis*) is generally made on a representative thin section of the rock, either by using a linear intercept method (Rosiwal analysis) or by point counting method. Chayes (1956) reviewed the methods employed by earlier workers as well as those proposed by himself and gives a detailed account of the statistical basis for the point count method. Reproducibility of the results obtained by this method and information regarding the relationship between the area of the thin section needed, the number of points to be counted, the grain size of the rock (expressed as an identity change—I.C. number) and the accuracy of the modal analysis are also given. The method consists of traversing the thin section at a set grid interval and making tallies of the different minerals observed at the grid points. The tally numbers for different minerals are recalculated to vol.% or wt.% (obtained by multiplying the tallies by the specific gravity of each mineral). Mineral identity can also be achieved by use of an automated electron probe microanalyzer. A computer can be programmed to set up a grid pattern for use by the microanalyzer.

Optical properties of minerals

Examination of suitably prepared thin sections (approximately 0.03 mm in thickness) of rocks or mounts of powdered minerals under a polarizing microscope is standard petrographic procedure. In rock thin sections, besides textural relationships, the identity of most

minerals can be established using their optical properties, such as refractive index (R.I.), relief, extinction angle, optically positive or negative character, and optic axial angle. Heavy-mineral analysis, the study of heavy minerals present in sedimentary rocks, is used in sedimentary petrography.

R.I. determination is best achieved by the "immersion method." A suitably oriented mineral grain is examined in oil mounts of varying R.I. values. The contrast in the R.I. values between the mineral and a given oil is tested using the "Becke line" effect or some other suitable method. When the R.I. of the mineral and oil match, the Becke line is invisible. The relief shown by a mineral is dependent on the difference between the R.I. values of the mineral and of the surrounding medium. In most cases the medium in which minerals or rocks are mounted is canada balsam (R.I. ~1.54). Extinction angle is the angle between a morphological direction, such as a cleavage or crystal edge, and one of the privileged optical directions in the mineral. Interference colors of minerals observed under crossed polarizers vary according to the birefringence, thickness, and orientation of a mineral. The order of the interference color may be determined by using a quartz wedge or Berek compensator.

Optically, minerals are classified into three groups: isotropic, uniaxial, and biaxial. In isotropic minerals the velocity of propagation of light is the same in all directions and hence these minerals have only one R.I. value. In uniaxial minerals light is split into two rays, the ordinary ray (O) and the extraordinary ray (E) each with its own velocity and associated R.I. value (ω and ϵ , respectively). They contain one optic axis. Uniaxial minerals can be either optically positive or negative. In biaxial minerals light is split into three rays (X, Y, and Z) and consequently they have three refractive indices (α , β , and γ , respectively); ω in uniaxial and β in biaxial minerals are considered diagnostic and are determined on grains suitably oriented (perpendicular to the optic axis) using the immersion method. The biaxial minerals are also subdivided into optically positive and optically negative varieties. These minerals contain two optic axes. The angle between the two is known as the optic axial angle or 2V. A qualitative estimate of the 2V value may be made by observing the configuration of the interference figure on mineral grains under conoscopic conditions.

Procedures for the determination of the optical properties of minerals are given in many books (e.g., Wahlstrom, 1969) as is a detailed listing of the optical properties of different

rock-forming minerals (e.g., Phillips and Griffen, 1981).

For accurate measurement of 2V values as well as other optical properties, a rotation apparatus or a universal stage, attached to the polarizing microscope is required. The simple rotation apparatus such as the spindle stage contains one or two axes of rotation and the universal or Federov stages employ four or five axes of rotation that enable proper orientation of a given mineral grain for accurate determination of several optical properties. The 4-axis stage contains two mutually perpendicular horizontal axes of rotation along with two vertical rotation axes, one of which is exclusively used for centering purposes. Optic axes can be located by proper tilting of the mineral grain on the different axes and the 2V value determined. Any of the various optical or morphological directions may be located precisely and their angular relationships measured either directly or by using a Wulff's stereographic net (Emmons, 1943). The universal stage methods are also used for the determination of twinning laws in plagioclases and other minerals, as well as for petrofabric analysis.

Reflected-light microscopy is used to study opaque ore minerals. Properties such as reflectivity, microindentation hardness, results of etch tests, and polarization properties are used in identification (e.g., Craig and Vaughan, 1981); tables for microscopic identification of ore minerals and their habits are given by Ramdhor (1968) and Uytendogaardt (1971) amongst others.

CHERUKUPALLI E. NEHRU

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Cross-references: *Petrochemical calculations*; *Petrographic microscope*.

PETROGRAPHIC MICROSCOPE

The petrographic, or polarizing, microscope, a compound instrument using plane-polarized transmitted light, has been applied for over a hundred years as a standard tool in the examination of rock textures and optical properties of minerals. Textures and most optical properties are normally observed in thin sections about 30 μm thick mounted in balsam or synthetic resin, but minerals may also be examined as individual grains immersed in liquids of known refractive index. Microscopes using reflected light are also available for the examination of opaque minerals. Polarization of light in modern microscopes is achieved by a synthetic material made of a plastic film containing a mass of fine-grained prismatic crystals of herapathite (a complex sulfate and iodide of quinine) with a parallel orientation. Early instruments used nicol prisms made of two elements of calcite cleavage rhombs cemented together in a particular optical orientation by balsam to achieve the same effect.

The lower polarizer or "polar" (Fig. 1) polarizes light from a conventional or monochromatic source in a plane parallel to the microscope axis and usually oriented E-W with reference to the field of view. The plane-polarized light passes through a system of iris diaphragms and condenser lenses to the thin section mounted on a 360° rotatable stage. During passage through the mineral lattice, the

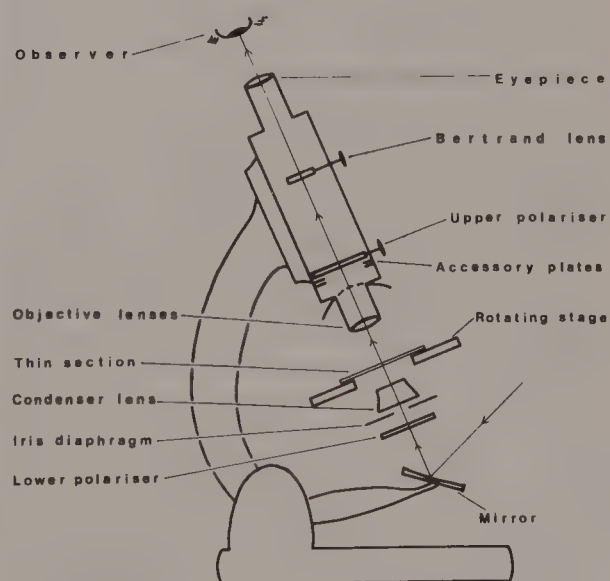


FIGURE 1. A petrographic microscope.

light is refracted and repolarized to a degree depending on the optical properties of the mineral species. It goes on through the objective lens system and thence through the eyepiece or ocular to the observer. Observation of minerals under such conditions with only the lower polarizer in position is referred to as plane-polarized light (PPL) or ordinary light (OL), although strictly the latter term should refer to observation when the lower polarizer is omitted. Properties noted in plane-polarized light include color and pleochroism (a function of absorption of light for different vibration directions), form, relief and relative refractive index (Becke test), cleavage, and alteration.

The upper polarizer (analyzer) polarizes light in a plane perpendicular to that of light polarized by the lower polarizer, i.e., usually N-S with reference to the field of view. With the analyzer in position, a condition referred to as crossed polars or "crossed nicols" (XN), optical phenomena are observed that result from differences in values of refractive indices, in effect different velocities of light, for particular directions in the mineral. Properties noted with crossed polars include birefringence, twinning, and extinction angle, the last providing a measure of the angle between crystallographic axes and principal vibration directions. A standard accessory for observation of properties with crossed polars is the Bertrand lens, which is a device inserted between the eyepiece and an analyzer to bring the image of interference figures into the focal plane of the eyepiece. The condenser lens beneath the microscope stage must also be in position at the same time in order to provide a convergent light beam.

Other attachments used in conjunction with the standard petrographic microscope include the universal stage, a device screwed to the microscope stage and which in modern versions provides five additional axes of rotation independent of the microscope axis. Tint plates of mica or gypsum and thin wedges of quartz mounted in carriers form additional standard accessories; the plates and wedges are inserted between the objective lens and analyzer in order to ascertain such properties as the sign of interference figures or the orientation of vibration directions relative to crystallographic directions.

BRIAN CHADWICK

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PETROGRAPHIC PROVINCE

It has been known for a long time that certain types of igneous rocks are present in characteristic associations that display not only a community of mineralogical and geochemical features but also continuity in relation to variations shown. For such associations of mineralogically and/or geochemically "consanguineous" rocks (rock associations), the term *petrographic province* was proposed by Judd (1886). To emphasize the origin of these associated rocks from a common primary source, Washington (1906) introduced the term *comagmatic region*.

Harker (1909) applied the term petrographic province to large regions of the Earth and distinguished the Pacific province (calc-alkaline suites) and the Atlantic province (alkali suites). However, initially it was not sufficiently emphasized that this broad twofold grouping was only roughly valid for volcanic rocks and those predominantly of Cenozoic age.

The terms Pacific suite and Atlantic suite were used by Niggli (1923) in the same way as Harker; in addition, the term Mediterranean suite was used for a K-rich suite that includes the Roman province with its abundance of leucite-bearing rocks. Other examples of igneous associations showing chemical and mineralogical consanguinity (named from geographical regions) are the (1) Thulean (Brito-Arctic—British Tertiary Volcanic) Province characterized by the effusion of tholeiitic magmas, and with the basalts and associated igneous rocks referred to as the Arctic suite and being intermediate in composition between rocks of the Pacific and Atlantic suites; (2) the Cascade Province of northwestern U.S.A., characterized by the products of calc-alkaline volcanicity. With these associations being dominated by volcanic products, petrographic provinces were commonly referred to as *volcanic provinces*.

Gradually as information about igneous rocks accumulated and understanding about their petrogenesis increased, the definition of petrographic province became vague and confusing so that its use has become largely obsolete (e.g., Turner and Verhoogen, 1951; Barth, 1952). Nevertheless, from a historical point of view the introduction of the concept of a petrographic (or volcanic) province was an early and major step towards a better understanding of the regularities in the generation of magmas. In

many ways it was the forerunner of the recognition of the importance of tectonic setting in igneous petrogenesis.

H. KLÁPOVÁ

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Cross-references: *British Tertiary Volcanic Province*; *Circum-Pacific plutonic and volcanic activity*; *Flood basalt*; *Igneous rocks—tectonic setting*; *Mediterranean suite*; *Mid-ocean ridge and ocean-floor petrology*; *Ocean-island igneous rocks*; *Pacific and Atlantic suites*; *Rift valley volcanism*.

PETROLOGICAL TERMS

The following are terms related to igneous and metamorphic petrology that are not set out in other articles in this volume.

Akyrosome—a subsidiary part of a complex rock, e.g., a vein in a migmatite.

Allocthonous—refers to a magma, and particularly to the neosome of a migmatite, derived from an extraneous source; the common use of the term is in structural geology.

Antistress mineral—a mineral whose formation in metamorphic rocks is favored by thermal activity and hydrostatic pressure rather than shearing stress, e.g., andalusite, forsterite.

Authigenic—refers to those constituents that were developed with or after the formation of the rock of which they are a part.

Autocthonous, **paraautocthonous**—terms that refer, respectively, to rocks that have been formed in situ (e.g., autocthonous granite) and to mobilized vein material (e.g., some neosome of migmatite) that has moved only a short distance and is characterized by variably diffuse to sharp margins.

Color index (CI)—the ratio of mafic to felsic minerals that expresses the leucocratic (color index 0–30), mesocratic (30–60), and melanocratic (60–100) character of a rock; used particularly for igneous rocks.

Crystalline (rock)—refers to a rock made up entirely of crystals, or fragments of crystals such as igneous rocks containing no glass and

metamorphic rocks resulting from recrystallization.

Dilation—deformation by change in volume, but not shape, e.g., a dyke intruded into a fracture causing walls of the fracture to move apart.

Essential mineral—one that is necessary for the classification or naming of a rock, but that is not necessarily present in large proportions.

Fabric—rock texture that combines crystallinity and grain size with shapes and arrangement of the constituents.

Granomerite—a holocrystalline rock in which the individual components can be distinguished without the use of a microscope.

Granularity—one of the components of the texture of a crystalline rock representing the size and grain-size distribution of constituent particles that contribute to a rock being granular.

Hard rock—a general term for igneous and metamorphic rocks, as distinguished from sedimentary rock (soft rock).

Homophanous—relates to rocks that not only show the same appearance (homogeneous nature) throughout but also show a massive plutonic structure, as in the case of large masses of neosome in migmatite complexes.

Hydatopneumatogenic—refers to a rock or mineral deposit formed by the action of both aqueous and gaseous agents.

Hysterogetic—refers to the final crystallization products of a magma.

Kyriosome—the framework of a complex rock, e.g., the major part of a migmatite.

Leucocratic—applied to light-colored rocks due to an abundance of felsic minerals; color index 0–30.

Lithology—Greek *lithos*, stone; the physical character of a rock; similar to or synonymous with petrography.

Macrocrystalline—refers to a rock or rock texture in which individual minerals are visible to the unaided eye or with the use of a hand lens.

Massive—homogeneous without layering, foliation, etc.

Megascopic—refers to observations made on the characters of rocks and minerals using the unaided eye or a hand lens.

Melanocratic—applied to dark-colored rocks due to an abundance of mafic minerals; color index 60–100.

Mesocratic—applied to rocks composed of approximately equal proportions of light-colored and dark-colored minerals; color index 30–60.

Mesocrystalline—refers to a rock or rock texture in which the sizes of individual crystals are between those of a macrocrystalline and a microcrystalline rock.

Microcrystalline—refers to a rock or rock texture (particularly in a groundmass) in which individual crystals can only be seen under the microscope.

Paragenesis—refers to the association of minerals in rocks and rock suites considered in relation to their origin, e.g., the sequential crystallization of minerals from a magma or the progressive development of metamorphic mineral assemblages.

Pneumotectic—refers to products and processes of magmatic crystallization affected to some degree by gaseous constituents of the magma.

Rheology—the study of deformation and flow of matter.

Selvage—a marginal zone of a rock mass that has distinctive features of composition and/or fabric, e.g., the margins of pillows in pillow lavas and the chilled borders (margins) of igneous intrusions.

Siliceous—refers to a rock with abundant SiO_2 , particularly free SiO_2 rather than as silicates.

Stress mineral—a mineral whose formation in metamorphic rocks is favored by shearing stress, e.g., chloritoid, kyanite.

System—a conceptual range of composition defined by a set of components in terms of which all compositions of the system can be expressed, e.g., MgO-CaO-SiO_2 .

Texture—the general physical appearance of a rock as shown by the crystallinity, granularity and fabric (mutual relationships) of constituent elements. There is no strict boundary between texture and *structure*; the latter is generally used for the larger features of a rock but some features termed as textures by some authors are referred to as structures by other authors.

Thermodiffusion—diffusion of material in solution as the result of a temperature gradient.

Transfusion—a complex series of processes involving the reaction of emanations and other materials, particularly associated with the uprise of material from deep sources; the products range from essentially solid, metasomatically altered rock to completely liquid magma, e.g., in the development of granitic rocks.

Vitrification (vitrification)—formation of glassy or noncrystalline substance.

D. R. BOWES

PETROLOGY—BRANCHES

Petrology (originally “petralogy” from Greek *petra*, rock, and *logos*, discourse) deals with the nature, occurrence and development of rocks. The following are particular branches of the topic dealt with in this volume.

Experimental petrology—concerned with experiments designed to reproduce in the laboratory conditions of formation of minerals and rocks with the aim of elucidating the nature of petrological processes.

Igneous petrology—concerned with rocks that solidified from hot, molten, or partly molten material (Latin *ignis*, fire).

Metamorphic petrology—relates to rocks formed by the mineralogical and structural adjustment of solid rocks to physical and chemical conditions (imposed at depths below the surface zones of weathering) that differ from the conditions under which the rocks in question originated (from the Greek meaning “change of form”).

Petrochemistry—a branch of petrology dealing with the study of the chemical composition of rocks.

Petrogenesis—a branch of petrology dealing with the origin and formation of rocks.

Petrography—a branch of petrology dealing with the description and systematic classification of rocks.

Volcanology (vulcanology)—a branch of igneous petrology that deals with the processes by which magma and its gases rise into the crust and are extruded on to the Earth’s surface and into the atmosphere (*volcanism*), and with the nature of the rocks formed by these processes.

D. R. BOWES

PETROLOGY—HISTORY

Petrology is the science that deals with the origin of rocks. Since rocks are major components of the Earth and are themselves composed of aggregates of minerals, petrology has close links with geology and mineralogy. Through most of the 18th century petrology was a rather ill-defined adjunct of mineralogy with descriptions of minerals and rocks occurring together in “mineralogical” treatises. By the end of the century, however, petrology had become more independent as evidenced by the publication of essays on specifically petrologic topics (e.g., de Launay, 1786, see Loewinson-Lessing). On the other hand, petrology became increasingly integrated with palaeontology and stratigraphy into the new systematic science of geology. Although petrology (petralogy of Pinkerton, 1811) remains, to the present, a part of geology, its importance is shown by the existence of many research institutes and university chairs devoted specifically to its study.

The history of petrology (Loewinson-Lessing, 1954) can be considered as a history of ideas,

the development of which is stimulated by progress in related subjects and also by technological advance. Nevertheless, although the most significant early technological innovation was the application of the microscope to petrologic research (about 1870), many important petrologic theories have an earlier origin. The period following the application of the microscope was characterized by an intensive interest in descriptive petrography, but since 1890 chemical and experimental petrology have become relatively more important, largely as a result of improvements in technique. Such improvements have been especially notable in the last 25 years, which have seen a great increase in the amount of money and manpower committed to petrology. The main developments in petrology are summarized in Table 1.

Geologic aspects

The geologic approach to petrology has a history going back to the early development of geology itself. It is also of fundamental importance inasmuch as it provides meaningful samples for petrographic and chemical investigation and the information necessary to devise meaningful experiments. This approach is concerned with the relationships of rocks to geologic structure and with their distribution and mode of occurrence, the last of these being important in the division of rocks into three major groups: sedimentary (e.g., Steno, 1669), igneous (e.g., Hutton, 1788, 1795), and metamorphic (Lyell, 1833). It was also significant with regard to the distinction of contact and regional metamorphism (e.g., Daubree, 1860), the division of igneous rocks into volcanic, plutonic (e.g., Hutton and Lyell), and hypabyssal (Rosenbusch, 1877; Brögger, 1886), and the subsequent definition of various plutonic and hypabyssal intrusive forms.

Although the study of volcanic forms is important, investigation of volcanic rocks also proceeds in another way, viz., through application of the principle of actualism (that the present is the key to the past). Thus, the study of volcanic rocks has been greatly advanced by work on actual eruptions and latterly by the establishment of observatories that continually monitor events at active volcanoes, such as those of Hawaii.

The other major "geological" aspect of rocks is concerned with the distribution in space and time. Such studies operate on different scales. They include work on metamorphic areas, individual volcanic centers, intrusions, and igneous complexes. On a larger scale, geologic studies integrated with petrographic and chemical investigation have given rise to the concepts

of the petrographic province (Vogelsgang, 1872; Judd, 1886) and the igneous association (e.g., Kennedy, 1938). The first concept arises from the observation that in certain regions igneous rocks having related petrologic characteristics and hence probably a genetic relationship have been emplaced or erupted during the same general epoch of activity. In addition, petrographic provinces conform to a limited number of major types, the rock associations.

Before about 1960 all the rocks available for study came from the Earth, and sampling there was restricted to the continents and oceanic islands. Since then, the scope of petrology has widened to include the direct investigation of rocks from the Moon and from the terrestrial ocean floors. Work on ocean-floor samples together with geophysical work has led to the formulation of the concept of sea-floor spreading and the plate tectonic theory. The development of the latter has stimulated renewed interest in what has long been a major concern of petrologists, i.e., the relationship of the origin of the rock associations to major tectonic processes like orogenesis (e.g., Suess, 1886; Gill, 1981).

Petrography

The petrographic approach overlaps with the geologic one in as much as it deals with field characteristics of rocks. Its more important aspect is, however, microscopic. Although a number of workers (notably H. C. Sorby and F. Zirkel) had previously used the microscope for petrologic study, the microscope did not really become an accepted research tool until Zirkel (1870), Rosenbusch (1877), and Fouqué and Michel-Lévy (1879) published their classic works on the microscopic properties of rocks and minerals. From 1870 to 1890, systematic petrography dominated petrology, but thereafter petrography can be seen to be just one of a number of approaches towards the elucidation of petrologic problems.

The main contribution of petrography to petrologic thought is probably in the field of classification (e.g., Johannsen, 1931–1938). Even with the great increase in the availability of chemical data, petrography remains the main basis for classification, whether through texture or mineral content and composition. Numerous schemes of classification for the igneous rocks have been proposed and several are still in use, but the system of Streckeisen (1976, 1979) is gaining increasingly widespread acceptance. In metamorphic petrology, petrography (especially texture) is again the main basis for classification and it has also provided—for example, through petrofabric analysis (e.g., Sander, 1930)—important petrogenetic information.

TABLE 1. Developments in petrology since 1770

Approximate period	Major trends and developments	Technical developments	Conceptual developments
1770–1870	Development of petrology as independent subject	Beginnings of experimental petrology (e.g., Hall, 1805)	Acidic and basic igneous rocks (e.g., Abich, 1841)
	Crude technology but many important ideas	Thin section making and early applications of the microscope (e.g., Sorby, 1858)	Primary magma (Scrope, 1825; Bunsen, 1851)
	Neptunist ideas discredited—recognition of metamorphic, sedimentary and igneous rocks; magmatic origin for the latter group established by the Plutonists (e.g., Hutton, 1788; Lyell, 1830–1833)	Classical analytical methods (e.g., gravimetric analysis), producing small amounts of data (e.g., Roth, 1861)	Magma mixing and liquid immiscibility as major processes of magma diversification (e.g., Durocher, 1857)
			Metamorphism (Lyell, 1833) Metasomatism (Naumann, 1849) Contact and regional metamorphism (e.g., Daubree, 1860)
1870–1890	The microscope becomes an accepted research tool	Common application of the microscope	Hypabyssal rocks (e.g., Rosenbusch, 1877; Brögger, 1886)
	Intense interest in petrographic description and classification (e.g., Zirkel, 1870; Rosenbusch, 1877; Fouqué and Michel-Lévy, 1879)	Beginning of Vogt's work on slags (1884)	Petrographic provinces (Vogelsgang, 1872; Judd, 1886) Contact metamorphic zones (Rosenbusch, 1877)
1890–1950	Decline in relative importance of petrography	Foundation of Geophysical Laboratory, Washington (1904)	Chemical classification of rocks (e.g., Cross et al., 1903)
	Chemical and experimental petrology become more important	Design of laboratory apparatus for experimental work at high temperature, pressure, and water vapor pressure	Magma types (Bailey et al., 1924)
		Classical analytical methods still used but early applications of spectroscopic methods	Fractional crystallization as major process in diversification of igneous rocks (e.g., Bowen, 1928)
		Petrofabric analysis (e.g., Sander, 1930)	Igneous rock associations (e.g., Kennedy, 1938)
			Regional metamorphic zones (Barrow, 1893) Facies concept (Eskola, 1915)
1950–1987	Expansion in manpower ^a and money committed to petrology	Experimental apparatus giving very high pressures	Importance of partial melting processes in petrogenesis of igneous rocks (Wyllie, 1977)
	Major technological breakthroughs in analytical and experimental petrology	Common application of rapid automated analytical methods giving great increase in analytical data ^b	Facies series and metamorphic belts (Miyashiro, 1961)
	Investigation of rocks of the ocean floors	Investigation of thermodynamics and kinetics of many petrologic reactions (Newton et al., 1981)	Relationship of origin and distribution of igneous and metamorphic rocks to plate tectonic activity
	Investigation of rocks from the Moon		

^a For example, between the wars, the British universities produced a yearly average of about 30 graduates in geology compared with about 1,200 graduates in 1984.

^b Chayes and Le Maitre (1972) estimate that there were ten times as many published major-element analyses in existence in 1972 as at the time of Washington's compilation in 1913. The increase in trace-element and isotope data is even more dramatic.

Chemical petrology

Progress in chemical petrology (petrochemistry) has been greatly affected by advances in analytical techniques. Early analytical methods were time consuming and expensive, so that the amount of analytical data obtained was relatively small. The last 25 years, on the other hand, have seen the development of sophisticated, rapid techniques, such as X-ray fluorescence, that have greatly increased the range of elements and isotopes that can be analyzed as well as the total number of analyses.

Analytical data have various applications to petrologic problems. They have been used as fingerprints of particular rock types (Roth, 1861; Washington, 1917) and in raw or processed form for classification (e.g., the norm classification of Cross et al., 1903, and Niggli, 1931). Analytical data also provide the basis for concepts like primary magma (Scrope, 1825), magma type (Bailey et al., 1924), and metasomatism (Naumann, 1849). Another important application is in petrogenesis, where isotope and trace-element studies have recently provided a promising way of assessing the relative contributions from various sources (e.g., upper mantle, crust) to petrogenetic processes (e.g., Maaløe, 1985).

Reactions involving rocks, minerals, and magmas can also be studied through the application of physical-chemical principles. Historically the main approach has been through the experimental determination of phase relations in silicate systems (see below), but thermodynamic and kinetic investigations of relevant reactions are making an increasingly important contribution to the understanding of petrogenesis.

Experimental petrology

Experimentation has been applied to many aspects of petrology, such as the structure and deformation of rocks (e.g., Daubree, 1860, 1879) and the mechanics of intrusion (e.g., Reyer, 1892), but its most important contribution has possibly been to petrogenesis through investigations of simple silicate systems and of the melting relations of rocks and minerals. The latter approach goes back to early in the history of petrology and in particular to Sir James Hall (e.g., 1805) who produced important results using primitive equipment such as a gun barrel. Subsequently, important information was obtained from industrial processes, particularly smelting (Vogt, 1884–1923), but the real breakthroughs were the development of small-scale laboratory apparatus capable of withstanding high temperature and pressure and the establishment of research institutes concentrat-

ing on experimental petrology, notably the Geophysical Laboratory, Washington (1904). In the 20th century, experimental petrology has provided a great stimulus to petrologic thought on diverse topics such as the petrogenesis of the igneous rocks (e.g., Bowen, 1928; Yoder and Tilley, 1962; Wyllie, 1977) and the stability ranges of metamorphic mineral assemblages (e.g., Saxena, 1983). Improvements in technique now allow systems investigated to approach more nearly the complexity of natural systems, but many of the conclusions of the early workers remain valid.

Petrologic theory

Experimental, chemical, petrographic, and geologic data provide the basis for petrologic theory. Although such theory has evolved in response to new techniques and new data, the main subjects of concern are few and have remained fairly constant throughout the development of petrology.

In igneous petrology, one of the earliest and most fundamental debates was concerned with the nature and origin of igneous rocks. Whereas the origin of certain rocks by consolidation of molten material erupted from volcanoes was known from antiquity, a “magmatic” origin for most “igneous” rocks was not even proposed before the mid 18th century. Before that, all rocks, including granite and basalt, were considered to be produced by precipitation from an aqueous fluid (Neptunist school of Werner, etc.), the only exceptions being the lava of active volcanoes. Gradually, however, men such as Desmarest (1771) and Faujar de Saint-Fond (1778) were able to prove the identity of ancient lavas (e.g., from the Auvergne) with those produced during modern eruptions, while Hutton (1795) and others extended the magmatic origin to certain granites and eventually also to other plutonic rocks.

Subsequently, much attention has been given to the nature and origin of magma. The question of a primary magma from which all igneous rocks are ultimately derived was first raised by Scrope (1825) and later by Bunsen (1851), who suggested two primary magmas, one acidic, the other basic. Regarding the availability of magma, there have been two main views: (1) that magma is present in discrete liquid layers within the Earth (Bunsen), or (2) that magma becomes available through the melting of solid silicate parts of the Earth (e.g., Reyer, 1877). More recent work, particularly high-pressure experimental work, has refined these ideas so that magmas are now thought to originate by melting or partial melting with variation in the composition of the

source rock or the conditions of melting giving rise to magmas of varying composition. Thus, basaltic magma is probably derived by dry partial melting of upper mantle peridotite, whereas andesitic compositions may be produced from the same peridotite by wet partial melting. Melting or partial melting of lower crustal material on the other hand, will probably give rise to more acidic (granitic) magma.

Three main processes have been considered important in the diversification of magma-mixing of magmas, assimilation, and differentiation. Durocher (1857) considered that the variety of igneous rocks could be produced by mixing of various proportions of Bunsen's primary magmas, but mixing is not now thought to be very important. Similarly, assimilation is now thought to be less important than suggested by some earlier authors (e.g., Daly, 1914). Differentiation embodies several processes, including liquid immiscibility, gas transfer, and fractional crystallization, the effect of which may be enhanced by gravity settling (e.g., Darwin, 1844), flotation, or filter pressing. Durocher (1857) thought that liquid immiscibility might be significant, but experimental work suggests that this is unlikely except in silicate-sulfide systems or in silicate systems of unusual compositions. The experimental results such as those of Bowen (e.g., 1928) further suggested that fractional crystallization might be the major process in differentiation, but more recently there has been a growing tendency to explain much of the diversity of igneous rock by variation in the melting processes that give rise to magmas rather than by the subsequent fractionation of a single (basaltic) parent magma.

The idea of metamorphism is implicit in Hutton's work, although the term was first proposed by Lyell (1833). The distinction between contact and regional metamorphism was established during the 19th century, largely through the work of Durocher, Delesse, and Daubree, while the concept of metasomatism derives from Naumann (1849).

The study of metasomatic reactions has subsequently given rise to important concepts such as migmatization (Sederholm, 1907) and granitization (Wegmann, 1938) but the major preoccupation of metamorphic petrologists has probably been with metamorphic mineral assemblages. The concept of metamorphic zones was introduced by Rosenbusch (1877) and Barrow (1893) who, in the Scottish Highlands, mapped a series of regional metamorphic zones, characterized by the appearance of "index minerals." The outstanding advance in the twentieth century is probably the development of the facies concept (Eskola, 1915). Following

Goldschmidt's (1911) application of the phase rule to metamorphism, Eskola correlated individual facies with broad fields of temperature and pressure, and experimentation and thermodynamics have since been used for the more detailed elaboration of the relationship between facies and conditions of metamorphism. Work on the field occurrence and associations of facies has led to the concepts of metamorphic facies series and metamorphic belts, e.g., (Miyashiro, 1961), which in turn have provided useful a framework for discussion of the relationship of metamorphism to major tectonic processes like orogeny and plate tectonic activity.

W. J. REA

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* Gives details of earlier works on the history of petrology, including many referred to here.

PHACOLITH

The name phacolite (Greek, *phakos*, lentil, and *lithos*, stone) was given by Harker (1909) to intrusive bodies whose "situation, habit, magnitude, and form . . . are all determined by the

circumstances of the folding." The type body is a dolerite mass in Ordovician strata of Shropshire, England.

Phacolith has been used instead of "phacolite" in most subsequent literature. The term *harpolith*, introduced by Cloos (1921) for a nearly vertically plunging, sickle-shaped, concordant igneous mass in Silesia, Czechoslovakia, is now recognized as essentially synonymous to the term *phacolith*, although some authors consider that as well as magmatic injection into strata being deformed, there was also subsequent deformation.

Phacolith, as a designation, is not only descriptive of a shape and structural relationships but also implies a mode of emplacement. Because of their restriction to crests and troughs of folds, phacoliths have roughly concavo-convex lenticular shapes and are concordant to their surrounding country rock. This is also the case for *saddle reefs*, such as the gold-bearing veins of Victoria, Australia. To prove that a body is a phacolith, it must be demonstrated not only that the mass has such a shape and structural relationships, but also that it resulted from consolidation of magma that was intruded during the tectonism responsible for the folding of the surrounding country rock. Without such proof, several alternatives frequently remain as possibilities, e.g., folded sediments that have undergone granitization or anatexis, folded sills, folded laccoliths, and some concordant igneous masses that have intruded previously folded rocks.

Intrusions that appear to fit the requirements set forth for phacolithic masses have been

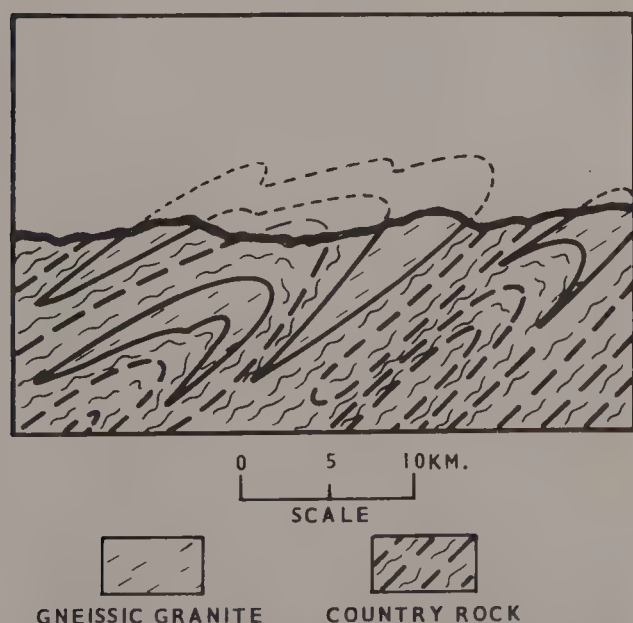


FIGURE 1. Phacoliths in cross-section. (After Buddington, 1929.)

recorded from many parts of the world (Buddington, 1929; Dietrich, 1957). Some especially noteworthy examples occur in the Sulitelma district, Norway, in SW Africa, and in the Grenville Province, Canada and adjacent northwestern Adirondack Mtns. of New York State, U.S.A. The phacolithic origin of those reported to occur in the northwestern Adirondacks, however, has been questioned; Dietrich (1963) and Engel and Engel (1963) suggest that these masses may, instead, represent sedimentary lenses that have undergone folding and partial mobilization, probably in response to anatexis, during regional tectonism.

Since its introduction, the term *phacolith* has been extended by many petrologists also to apply in a generic sense to a particular mode of emplacement for magma. Phacolithic emplacement is conceived to involve syntectonic (synkinematic), permissive (permitted) intrusion, i.e., to involve flowage into real or potential cavities created by tectonic stress. Permissive intrusion—as distinguished from forcible (forceful) intrusion—was first defined by Pirsson (1914) as intrusion in which magma is emplaced by flowing into space opened for it by some force other than that exerted by the invading magma itself.

R. V. DIETRICH

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Cross-references: *Emplacement—modes; Igneous intrusions—structural behavior; Intrusion.*

PHASE EQUILIBRIA—See Vol. IVA

PHASE RULE

The *phase rule*, first proposed by J. Willard Gibbs, relates the number of degrees of

freedom (F) to the number of phases (P) that may be in equilibrium in a system containing C components:

$$F = C - P + 2.$$

The number of degrees of freedom is the number of variables that may be arbitrarily specified in describing the equilibrium state of the system. For example, in the case of water in equilibrium with ice there is a system of two phases and one component. The number of degrees of freedom, F , is 1, and the system is referred to as univariant. If the pressure of the system is specified, the temperature is fixed by the constraint of equilibrium between water and ice.

When applying the phase rule to systems containing a number of chemical species, care must be taken to specify correctly the number of components. Where chemical reactions are possible, the number of components is the minimum number required to generate the system. For the system albite + jadeite + quartz, there are only two components, since



The number of degrees of freedom for equilibrium is 1 for coexistence of all three phases.

Conclusions drawn from use of the phase rule must necessarily be qualitative. However, in geologic situations they may provide useful insights. In geologic systems, univariant ($F = 1$) and invariant ($F = 0$) equilibrium assemblages are rare. This led Goldschmidt to propose the mineralogical phase rule. By setting $F = 2$, he obtained the relationship that the number of phases possible at equilibrium is equal to the number of components. The presence of phases in excess of the number of components proves that either the special P - T requirements of the invariant or univariant equilibria obtained, or that the phases are a disequilibrium assemblage.

Korzhinskii (1949) proposed a modified phase rule for open systems:

$$P = C - M$$

where M is the number of perfectly mobile components. This formulation also assumes divariant equilibria. The main requirement for a perfectly mobile component is that its chemical potential be fixed by circumstances outside the system under consideration. Thompson (1970) has noted that a phase rule similar to that of Korzhinskii may be applied to mineralogical systems in which only a portion of the total assemblage is of interest.

V. M. OVERSBY

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PHASE TRANSFORMATION

A phase is a physically and chemically distinct and homogeneous portion of a thermodynamic system. Phase transformation, in a rigorous sense, is the conversion of one phase to another that is physically different from the original. The bulk chemical composition of the phase, however, remains unaltered after the transformation. Such a transformation may be induced by a change in temperature and/or pressure of the system. An example is the change in state of a single-component system, such as that which occurs in the vaporization and melting of ice (solid H_2O). A phase transformation between solid phases is often called a *polymorphic* transformation; each phase is called a *polymorph*. For example, graphite and diamond are the polymorphs of carbon, while quartz, tridymite, cristobalite, coesite, and stishovite are all polymorphs of SiO_2 .

Phase transformation, in a broad sense, refers to the conversion of one assemblage of phases in a heterogeneous system to another, without an alteration in the bulk chemical composition of the system. Such a usage of the term phase transformation is common in petrologic literature (e.g., Schairer, 1951). For instance, the change from the gabbroic assemblage (plagioclase + ferromagnesian pyroxenes) to the eclogitic assemblage (jadeite + garnet) is often referred to as a phase transformation. Although the bulk chemical composition of the system is unaltered in this case, the chemical composition of each mineral phase is changed before and after the transformation of the assemblage. Thus, this phenomenon fails to obey the rigorous definition of the term phase transformation, and should be referred to more appropriately as a chemical reaction of the phases.

Thermodynamics of phase transformation

A system in thermodynamic equilibrium at given temperature and pressure conditions should be composed of the phase (or phases) that would yield the minimum Gibbs free energy for the system. When these conditions are altered and another phase yields a lower Gibbs free energy for the system than the

original, a transformation to the new phase should take place. The Gibbs free energy of the system thus changes continuously and at varying rate as the pressure and temperature of the system are changed across the phase transformation. In thermodynamic terms, phase transformations can be classified according to the behavior of the Gibbs free energy in the vicinity of the conditions under which the transformation occurs. When volume and/or entropy, which are both first partial derivatives of the Gibbs free energy, undergo a discontinuous change at the phase transformation, the change is called a first-order transformation (Fig. 1). If the first derivatives change continuously but the second derivatives, such as specific heat and compressibility, undergo a discontinuous change, the change is called a second-order transformation (Fig. 2).

The effect of temperature (T) and pressure (P) on a first-order phase transformation in which the Gibbs free energies of the original and new phases become equal can be expressed by the Clausius–Clapeyron equation

$$\frac{dP}{dT} = \frac{S_2 - S_1}{V_2 - V_1}$$

where S and V denote entropy and volume, respectively, and the subscripts indicate the respective phases. A discontinuous change in volume and entropy at a first-order phase transformation indicates that a sudden transition in interatomic (or intermolecular) distances and/or in the degree of regularity in the atomic

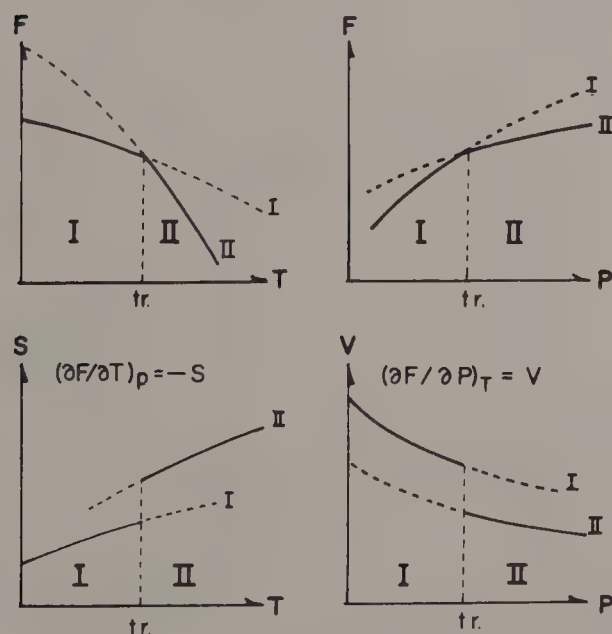


FIGURE 1. Changes of the Gibbs free energy (F), entropy (S) and volume (V) as a function of temperature (T) and pressure (P) across a first-order phase transformation.

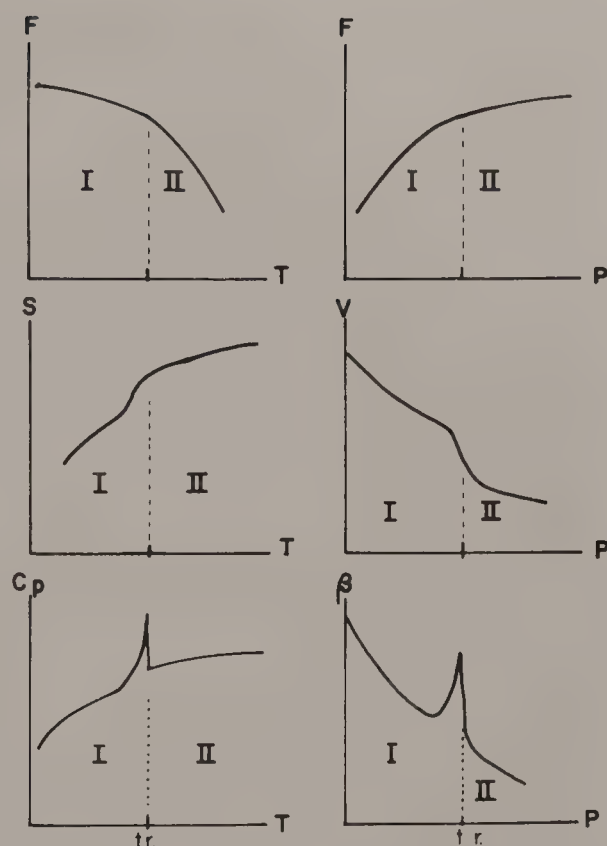


FIGURE 2. Changes of the Gibbs free energy (F), entropy (S), volume (V), specific heat (C_p) and compressibility (β) as a function of temperature (T) and pressure (P) across a second-order phase transformation; the specific heat $(\partial S/\partial T)_P$ and compressibility $(1/V)(\partial V/\partial P)_T$ are both the second derivatives of the Gibbs free energy.

(or molecular) arrangements has occurred. Accordingly, typical first-order transformations are the result of the following.

1. Changes from one physical state, gaseous, liquid, or solid, to another (e.g., water vapor–water–ice).
2. Polymorphic transitions accompanied by such major changes in atomic arrangement in the crystal as a change (i) in coordination number (e.g., α -iron (body-centered, cubic)– γ -iron (face-centered cubic), and coesite–stishovite); (ii) in symmetry (e.g., graphite–diamond, and quartz–tridymite–cristobalite), and/or (iii) in interatomic distance (e.g., cerium I (face-centered cubic)–cerium II (face-centered cubic)) (Adams and Davis, 1962).
3. Diffusionless (or martensitic) phase changes in multicomponent systems (e.g., austenite (face-centered cubic)–martensite (body-centered tetragonal) in the Fe–C system).

Typical second-order transformations are brought about by the following.

- Changes in the degree of ordering of the atoms in crystalline solids (e.g., the order-disorder transformation of albite ($\text{NaAlSi}_3\text{O}_8$), in which the distribution of the Al and Si atoms in tetrahedral sites becomes random at high temperatures).
- Polymorphic transitions accompanied by minor changes in atomic arrangement, such as a change in symmetry and bond angles that may be caused by lattice distortion (e.g., α -quartz- β -quartz).
- Changes in electronic behavior unaccompanied by changes in the atomic arrangement: these changes cause the discontinuous variation of such physical properties as magnetic, electric, dielectric, and optical properties (e.g., the Curie point ferromagnetic transformation and the Néel temperature antiferromagnetic transformation).

Geologic implications

In petrology, studies of the polymorphs of rock-forming minerals can provide information about the temperature and pressure conditions that existed at the time of formation of the rocks. Studies of andalusite-sillimanite-kyanite (polymorphs of Al_2SiO_5), calcite-aragonite (polymorphs of CaCO_3), and graphite-diamond (polymorphs of carbon), commonly serve as the basis for estimation of the temperature and depth at which the rocks were formed (Fyfe et al., 1958).

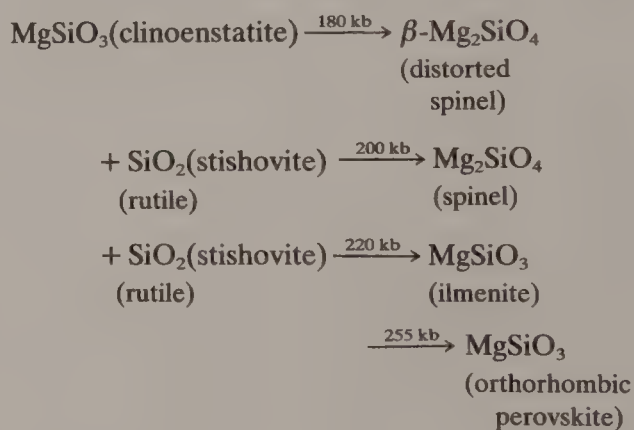
Since the formation of coesite and stishovite requires high-pressure conditions (Stishov and Popova, 1961) exceeding the pressures reached in the crustal environment, their occurrence in quartz-bearing sedimentary rocks suggests that the source of such extreme conditions is catastrophic. The association of these minerals with meteorite fragments in Canyon Diablo Meteor Crater, Arizona, indicates that they were formed during the impact of meteorites with the Earth (Chao et al., 1962) and the occurrence of these minerals has been accepted as a criterion for a meteoritic origin of the crater. On the basis of their occurrence in crater rocks, an impact origin of several other craters has been confirmed.

The occurrences of high-pressure polymorphs of olivine and pyroxene have been reported in Tenham and Coorara meteorites. The olivine polymorph that has a spinel structure has been named ringwoodite (Mason et al., 1968), and the pyroxene polymorph that has a garnet structure with some silicon atoms in the sixfold coordination has been named majorite (Smith and Mason, 1970). These high-pressure polymorphs are considered to be of extraterrestrial origin.

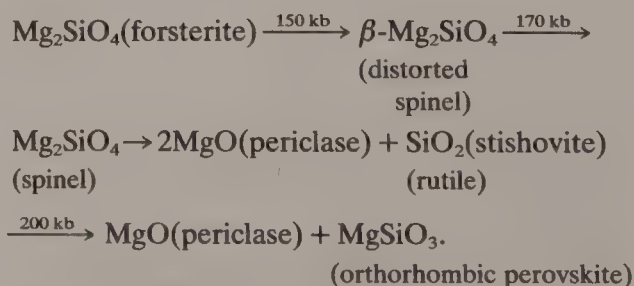
The high rate of increase in the seismic wave velocities with depth that occurs in the Earth's mantle at depths between 400 and 900 km has been interpreted by Birch (1952) as being a result of phase transformations of ferromagnesian silicate minerals such as olivines and pyroxenes. In such denser polymorphs, the silicon atoms are surrounded by six oxygen atoms (or are in sixfold coordination) rather than being surrounded by four oxygen atoms as found in less-dense ordinary rock-forming minerals. Closer proximity of silicon and oxygen atoms would result in a stiffer bonding between the atoms, and hence cause an increase in the speed of the elastic waves propagating through minerals.

Experiments under high-pressure and high-temperature conditions have shown that orthorhombic olivines transform to the cubic spinel phase (ringwoodite) (e.g., Ringwood, 1958) at high pressures. However elastic properties of the spinel phase of olivine (Anderson, 1967) indicate that the olivine-spinel phase transformation accounts partially for the observed increase in the seismic velocity in the transition zone, 400–900 km deep in the mantle.

Still denser phases of ferromagnesian silicates have been synthesized at high pressures (up to 300 kb) and at temperatures of 900–1,400°C (e.g., Ming and Bassett, 1975; Liu, 1976). The following sequence of phase transformations in MgSiO_3 and Mg_2SiO_4 has been observed at an estimated temperature of about 1,000°C:



and



The names in the parentheses after the

chemical symbols indicate mineral names, and those under the chemical symbols indicate type structures. Magnesian garnet (pyrope) has been found to be stable up to a pressure of 235 kb, and the denser polymorphs in the ilmenite and perovskite structures would become stable at greater pressures (Liu, 1977). These observations suggest that several phase transformations in silicate minerals appear to occur in the transition zone in the Earth's mantle.

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Cross-references: *Eclogite*; *Metamorphic petrology—use of thermodynamics*; *Phase rule*; *Reaction principle and reaction series*; *Shock metamorphism*; *Upper mantle—experimental petrology*.

PHENOCRYST

A *phenocryst* is a large crystal in an igneous rock. It is a cognate crystal, almost invariably early formed, and is genetically distinct from a

xenocryst (Iddings, 1889). Its derivation is from Greek *phaneros*, conspicuous, and *kristallos*, a crystal. Large igneous crystals are designated phenocrysts if they contrast sharply in size with the enclosing groundmass; this implies a size ratio of at least about 5:1. Porphyritic texture consists of phenocrysts in a groundmass and is typical of the rock porphyry.

Phenocrysts are free-growth (unimpeded) forms that have developed while supported by a fluid phase. Slowly grown crystals tend to be more nearly perfect, whereas rapidly grown crystals contain twins, zoning, mosaic structure, and inclusions of solid, liquid, or gas phases. Undisturbed conditions are favorable for the formation of crystal faces (to give euhedral crystals) even in minerals that do not otherwise tend to be well formed. Irregularities in shape result from rapid growth, magma-flow and mutual interference. Changes in physical conditions may lead to dissolution of the phenocryst, which may become embayed or corroded. Synneusis (mutual attraction of freely floating crystals) may result in doublets, twins or glomeroporphyritic texture.

Porphyritic texture is strongly dependent on nucleation and growth rates. The term megacryst is also used for large crystals but is nongenetic and is also applied to metamorphic crystals. Typical igneous megacrysts are interpreted by Binns et al. (1970) to be genetically distinct from phenocrysts.

ALAN SPRY

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Cross-references: *Hypabyssal rocks*; *Igneous rocks—classification and nomenclature*; *Porphyry*; *Textures of igneous rocks*; *Volcanic rocks*.

PHONOLITE

Phonolites are fine-grained, undersaturated, intermediate igneous rocks, with or without phenocrysts, consisting primarily of alkali feldspars and feldspathoids together with pyroxenes and amphiboles, usually sodic varieties. The name, due to M. H. Klaproth (see Johannsen, 1938), is an allusion to the ringing sound supposedly produced by striking them with a hammer. They are the extrusive equivalents of the feldspathoidal syenites. The common feldspathoid is nepheline (implied when the term “phonolite” is used unqualified), but leucite, nosean, haüyne, sodalite or analcime may be present and may be the only

feldspathoid; melilite may be present. Phonolites may be partly glassy and may show feldspathoid only in the norm. With nepheline > alkali feldspar they grade into *phonolitic nephelinite* (phonolitic foidite in the classification of Streckeisen, 1976), and when feldspathoid is <10% into *phonolitic trachytes*. With increasing plagioclase, phonolites grade (through tephritic phonolite and phonolitic tephrite) into tephrites and basanites (when olivine-bearing).

Clinkstone, *echodolite*, and *klingsstein* are archaic synonyms of phonolite.

Mineralogy

Feldspars, as groundmass (often showing trachytic texture) or as phenocrysts, may be members of the plagioclase–anorthoclase series or sodic sanidines. Two-feldspar (plagioclase + Na-sanidine) phonolites may occur. Nepheline may occur as a groundmass or phenocryst phase, sometimes as overgrowths on sanidine. Leucite, less common, is usually idiomorphic, both in the groundmass or as phenocrysts. Phase equilibria governing feldspar–nepheline or leucite relations are discussed by Hamilton and MacKenzie (1965). Other feldspathoids, such as nosean, haüyne, or analcime (in the rare blairmorites—see Woolley and Garson, 1970) may occur as primary euhedral phases or in the groundmass.

Aegirine-augite is the predominant mafic phase, often zoned from diopsidic cores to acmitic margins. Amphiboles range from somewhat sodic hornblende to riebeckite–arfvedsonite. Biotite is less common and often shows evidence of resorption. Titaniferous magnetite may occur as a primary phase or as a secondary product resulting from resorption of mafic phenocryst phases. Common accessories are apatite, zircon, rutile, and olivine.

Distribution and genesis

Like other alkaline rocks, phonolites characteristically occur in continental rift settings, such as E Africa (e.g., Woolley and Garson, 1970), or older rifts such as the ca. 1,200 Ma old Gardar rift province in S Greenland (Emeleus and Upton, 1976). In E Africa they occur together with normal calc-alkaline basic and acidic extrusive rocks and are also associated with central complexes often involving carbonatite. It is clear that there is close and persistent association of rifting and alkaline volcanicity in the continental regions and that the tectonic setting for igneous activity may profoundly affect the course of magma development. Modification of mantle rocks beneath rifts

by migrating CO₂-rich fluids has been suggested by many workers (reviewed in Bailey et al., 1980 and Fitton and Upton, 1987).

Phonolites are also repeatedly observed as minor members of the oceanic olivine basalt–trachyte association. Most phonolites have bulk compositions which, in terms of normative feldspars and feldspathoids, plot close to the liquidus minimum in “petrogeny’s residua system” (Hamilton and MacKenzie, 1965). This minimum may be reached by fractionation of alkali basalt melts via undersaturated trachytes or by fractionation of magma near nephelinite in composition. Nephelinization of lower crust by nephelinite magma and subsequent mobilization has also been postulated (Woolley and Garson, 1970; see **Feldspathoidal rocks—genesis**).

Other terms

Apachite—a nepheline phonolite rich in alkali-pyroxenes and amphiboles (Apache Mts., Texas, U.S.A.).

Blairmorite—an analcime phonolite with analcime phenocrysts in a groundmass of analcime, sanidine, and alkali pyroxene (Blairmore, Alberta, Canada).

Kenyte—an olivine-bearing phonolite containing anorthoclase, nepheline, acmite–augite, Na-amphibole, olivine, apatite, and opaque ores; the groundmass is trachytic or hyalopilitic (Mt. Kenya, Kenya).

Marienbergite—a plagioclase-bearing phonolite with natrolite instead of nepheline (Marienberg (Mariánské Lázně) Czechoslovakia).

Munionsgite—a hypabyssal rock similar to tinguaita containing nepheline, orthoclase, acmite, and sometimes cancrinite.

Murite—an olivine melaphonolite rich in nepheline.

Nephelinitoid phonolite—a general term for a phonolite in which feldspathoids are more abundant than feldspars.

Tinguaita—a textural variety of phonolite typically found in dykes; feldspar and nepheline phenocrysts are in a groundmass containing conspicuous acicular crystals of acmite arranged in a radial or criss-cross pattern (Tingua Mts., Brazil).

Trachytoid phonolite—a general term for varieties of nepheline phonolite containing preponderant alkali feldspars and showing trachytic texture.

Viterbite—a leucite phonolite with phenocrysts of sodic sanidine and leucite with smaller proportions of calcic plagioclase, augite, biotite, apatite, and opaque oxides (Viterbo, Italy).

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Cross-references: *Alkaline rocks—undersaturated; Feldspathoidal rocks—genesis; Intermediate igneous rocks; Tephrite, basanite; Volcanic rocks.*

PHYLLITE

Phyllite is a fine-grained, pelitic rock characterized by a penetrative parting, the surfaces of which display a lustrous sheen. Regarded as intermediate in metamorphic grade and grain size between slate and mica schist, phyllite is distinguished from slate by the sheen on parting (cleavage) surfaces, and from mica schist in that the mineral constituents of the schist are visible without the aid of a microscope. The term (from the Greek *phyllos*, leaf, a reference to its characteristic cleavage) was first used by C. F. Naumann in the late 1850s, but the original reference is lost.

The dominant mineral composition is the same as that of slate, with dimensionally aligned muscovite, chlorite, and quartz the major constituents. The most common minor constituents are iron ores (magnetite or pyrite) and feldspar clasts, with biotite sometimes present although generally where metamorphism reached this grade the rock is a schist.

Like slates, the major fabric element of phyllite is the penetrative parting, or slaty cleavage, caused by the nearly perfect alignment of the major minerals. This planar fabric is commonly undulatory as the result of superimposed folds or crenulations. Rarely, in

polyphase deformed terrane, relicts of it remain in porphyroblasts as indicators of changing *P–T* conditions during metamorphism.

Phyllite is regarded as part of a progressive metamorphic series of pelitic rocks beginning with the premetamorphic shale: the rocks may grade through the series slate–phyllite–mica schist with increasing degree of recrystallization. Like slate, it is commonly in the chlorite zone of the greenschist metamorphic facies but occasionally it contains biotite in the transition zone from chlorite to biotite grade.

In hand specimen or at outcrop, phyllite may be indistinguishable from *phyllonite* which is the product of retrograde metamorphism of coarser-grained and higher-grade rocks. As in phyllite, the recrystallized mica or chlorite define schistosity–penetrative parting surfaces but relicts of the protolith generally remain, including porphyroclasts and relict minerals. The term phyllonite was originated by Sander (1911).

INA B. ALTERMAN

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Cross-references: *Cataclastic rocks; Greenschist facies; Polyphase metamorphic mineral growth; Regional metamorphism; Retrograde metamorphism; Schist; Slate.*

PICRITE

G. Tschermak, in 1866, first applied the term *pikrit* to melanocratic derivatives of teschenitic igneous rocks from Silesia (on the border between Czechoslovakia and Poland), rich in olivine and containing hornblende, titaniferous augite, and ilmenite with small amounts of plagioclase and analcime and accessory biotite and apatite (see Johannsen, 1938). Owing to the early usage of the term carrying an implication of alkaline affinity it is common to find orthopyroxene-bearing varieties labeled with a suitable prefix, e.g., bronzite picrite. Modern usage of the term has been extended to cover igneous rocks of various kindreds, grain sizes, and textures with compositions roughly intermediate between olivine melagabbros (or basalts) and peridotites. A typical picrite, therefore, consists dominantly of olivine with

smaller proportions of clinopyroxene or orthopyroxene, or both, and with salic constituents (plagioclase $\pm$ feldspathoids) forming 5–25% of the mode. Hornblende, biotite, and ilmenite or titanomagnetite are common minor constituents, though chromite may be the dominant ore mineral in magnesia-rich picrites.

The term *oceanite* (picrite-basalt) has been used to describe picrites that have been derived directly from basalts through magmatic accumulation of olivine by crystal settling. Olivine-rich cumulates variously approximating to picrite are found in many large- and medium-sized differentiated, gabbroic intrusions, e.g., the Stillwater intrusion, U.S.A. (Wager and Brown, 1968). *Madeirite* is a porphyritic variety of alkali picrite (Madeira, Atlantic Ocean) and *schriesheimite* is a hornblende picrite (Schriesheim, W Germany).

Not all picritic rocks have originated from basalt through crystal settling of olivine, however, as evidenced by the occurrence of picrite as hypabyssal intrusions, e.g., in the British Tertiary Volcanic Province (see Wyllie, 1967). Many of these minor intrusions display wide variations in mineral proportions and textures resulting from the combined effects of flowage differentiation and crystal settling in the low-viscosity picritic magma. The parental magma of the Great Dyke of Zimbabwe is considered to have been picritic (Cawthorn and Davis, 1982) and fine-grained extrusive peridotitic and picritic lavas (komatiites) have been described from a number of Archaean greenstone belts.

J. TARNEY

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Cross-references: *Basalt*; *Gabbro* and *norite*; *Magmatic differentiation*; *Peridotite*; *Ultramafic lavas*.

PLUTONIC ROCKS

In the restricted sense, *plutonic rocks* are coarse-grained crystalline igneous rocks formed by consolidation of molten rock material or magma below the Earth's surface (Latin *Pluto*, god of the Underworld). In the 18th century the name "plutonist" was applied to those geologists, led by J. Hutton who regarded granite as a product of molten rock material formed at depth, in opposition to the "neptunists" led by A. G. Werner who regarded granite as a precipitate of the early ocean. Plutonic action (later modified to plutonic activity or *plutonism*) was the name given by Lyell (1875) to "all those modifying causes brought into play at great depths and under conditions never exemplified at the surface. To this plutonic action the fusion of granite itself in the bowels of the Earth as well as the development of the metamorphic texture in sedimentary strata may be attributed." Lyell recognized four classes of rocks—aqueous, volcanic, plutonic, and metamorphic. He grouped the last two together as the "primary rocks," both being the products of plutonic action.

In the late 19th century a threefold subdivision of igneous rocks based on the environment of crystallization as inferred from field relationships and texture came into use. The plutonic rocks, thought to have crystallized deep in the Earth's crust, were contrasted with the hypabyssal and volcanic rocks thought to have crystallized at shallow depths or at the surface, respectively. The term hypabyssal fell into disuse as it became apparent that the criteria used to distinguish rocks assigned to a near-surface environment were not primarily dependent on depth of intrusion. Some petrologists now refer igneous rocks to two broad categories, viz., the fine-grained or glassy volcanic rocks formed at or near the surface and the predominantly coarse-grained plutonic rocks formed at depth or in large magmatic bodies emplaced at shallow depths.

Plutonic and volcanic associations were established by Kennedy and Anderson (1938) with a distinction based on the characteristic mode of occurrence of the main families of igneous rocks. The *plutonic associations* were principally granitic or granodioritic (acidic) in composition, produced by remelting of the "granitic" layer of the crust. The *volcanic associations* were predominantly basic, effusive, and nonorogenic and were derived from the melting of deeper layers of the crust or mantle. Turner and Verhoogen (1951) classed as plutonic associations those igneous rock series whose members are predominantly plutonic, including not only granitic and granodioritic

assemblages but also certain series composed predominantly of basic and ultrabasic igneous rocks. Read (1957) united the plutonic association with the deep-seated migmatites and metamorphic rocks of the orogenic belts in a single *plutonic series* produced by the processes classified by Lyell under the term plutonic action. In this usage, plutonic rocks form the bulk of layers 2 and 3 of the ocean crust and the deep continental crust.

Plutonic rocks (plutonites) cover the compositional range from acidic through intermediate and basic to ultrabasic, with granite, granodiorite, syenite, diorite, gabbro and peridotite as examples (see **Plutonic rocks—classification and nomenclature**). They are formed by the crystallization of a silicate magma intruded into the crust. The thickness of such intrusions was seldom less than a few hundred meters or greater than 20 km. The relatively slow cooling of large bodies of magma allows time for complete crystallization to take place and for the component minerals to grow to such a size that they are recognizable to the unaided eye, i.e., they are phaneritic.

During the solidification of large intrusions, early-formed crystals may be segregated and residual liquids may be separated from them at various stages. Such processes of magmatic differentiation may lead to the formation of several rock types within a single intrusion and further chemical variations may result from contamination of the magma by material assimilated from the enclosing country rocks. The characteristic textures of plutonic rocks distinguish them from the rapidly-cooled volcanic rocks. Plutonic rocks are holocrystalline (without glass) and their average grain size is 1–5 mm. Porphyritic textures are shown by many granites. The grain-to-grain relationships reflect the history of crystallization. Minerals that crystallized at early stages may be idiomorphic (euhedral), whereas those formed at late stages tend to be granular (anhedral), irregular, or poikilitic. Most are subhedral and hypidiomorphic texture is commonly referred to as *granitic texture*.

The large intrusions in which most plutonic igneous rocks occur include batholiths, laccoliths, lopoliths, stocks, bosses, diapirs, and ring complexes. The general term *pluton* is commonly used with reference to such deep-seated masses of plutonic rock.

JANET V. WATSON

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Cross-references: *Aureole*; *Batholith*; *Diapirism*; *Emplacement—modes*; *Hypabyssal rocks*; *Igneous intrusions—structural behavior*; *Igneous rocks—classification and nomenclature*; *Intrusion*; *Magma*; *Plutonic rocks—classification and nomenclature*; *Textures of igneous rocks*; *Volcanic rocks*.

PLUTONIC ROCKS— CLASSIFICATION AND NOMENCLATURE

The nomenclature of igneous rocks in general has evolved in an ad hoc fashion over two millenia, with names being derived from localities, appearance of the dressed stone, and physical properties. Some common names, e.g., syenite and porphyry, date back to antiquity. Since its earliest days as a scientific discipline, petrology has witnessed attempts to impose some systematic rules on nomenclature. Ideally names should be rigorously defined (on paragenetic rather than petrogenetic criteria) and be informative, if not self-explanatory. Again, ideally, the terms should be defined on the basis of gross petrographic characteristics rather than subtle variations in whole-rock or mineral chemistry. Particularly confusing in this respect has been the practice of naming varieties after localities and giving new names to variants of established species, often when the criteria on which the variation has been established are cryptic in the extreme. Any systematic nomenclature must be internationally acceptable, although igneous petrology contains many rock names that mean very different things in different countries; the use of “quartz monzonite” in N America, Europe, and U.S.S.R. is one example and the French use of “granulite” for biotite granite is another.

Any internationally accepted systematic nomenclature must ultimately be a compromise; what follows here is a summary of the recommendations made by the IUGS Subcommittee on the Systematics of Igneous Rocks presented to the meeting of the International Geological Congress in Montreal, August 1972 (see Streckeisen, 1976; Le Maitre, 1989).

Basis for classification and nomenclature

These recommendations concern all “igneous or igneous-looking” rocks, the “massive Ges-

teine" in the sense of H. Rosenbusch, or rocks formed by "plutonic activity" in the sense of C. Lyell. These recommendations can therefore be

applied to plutonic rocks intruded as magma, orthogneisses, and such problematica as members of the charnockite suite. Names are based on the actual (i.e., modal) mineralogy of the rock, and the system attempts to fulfil three basic requirements: (1) correspondence with natural relationships, (2) acceptability to geoscientists worldwide, particularly with regard to historical tradition, and (3) simplicity and ease of usage.

The classification used here is based on the relative proportions of five mineral groups, initialled Q, A, P, F, and M (Figs. 1, 2).

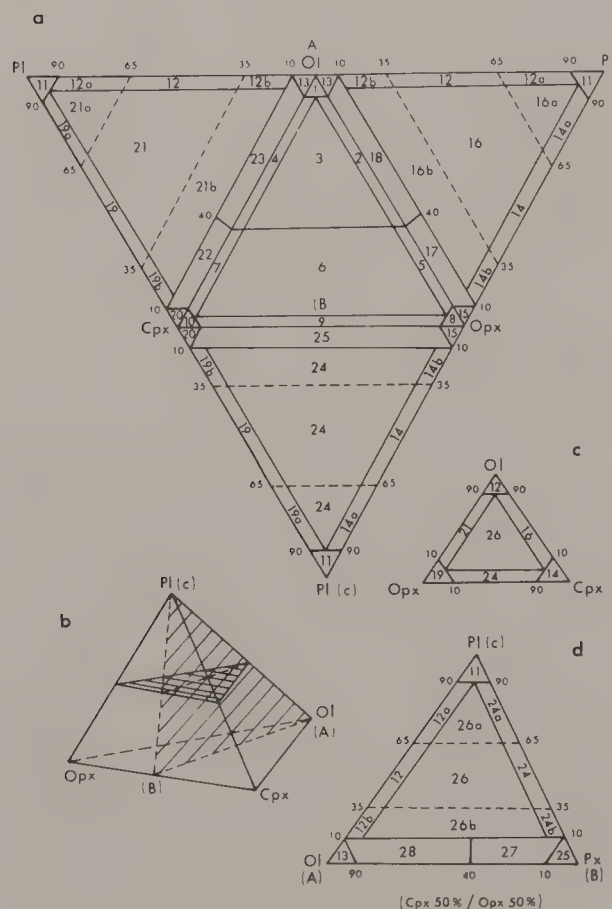


FIGURE 1. Ultramafic and gabbroic rocks (M + P), where M = olivine(ol) + orthopyroxene(opx) + clinopyroxene (cpx) and P is calcic plagioclase (pl).

Root names: 1, dunite (olivinite); 2, wehrlite; 3, lherzolite; 4, harzburgite; 5, olivine clinopyroxenite; 6, olivine websterite; 7, olivine orthopyroxenite; 8, clinopyroxenite (diopsidolite, diallagite, augitoidite, etc.); 9, websterite; 10, orthopyroxenite (enstatolite, bronzitolite, etc.); 11, anorthosite (plagioclase); 12, troctolite (a = leuco-, b = mela-); 13, plagioclase-bearing dunite; 14, gabbro (a = leuco-, b = mela-); 15, plagioclase-bearing clinopyroxenite; 16, olivine gabbro (a = leuco-, b = mela-); 17, plagioclase-bearing olivine clinopyroxenite; 18, plagioclase-bearing wehrlite; 19, norite (a = leuco-, b = mela-); 20, plagioclase-bearing orthopyroxenite; 21, olivine norite (a = leuco-, b = mela-); 22, plagioclase-bearing olivine orthopyroxenite; 23, plagioclase-bearing harzburgite; 24, gabbro-norite (a = leuco-, b = mela-); 25, plagioclase-bearing websterite; 26, olivine gabbro-norite (a = leuco-, b = mela-); 27, plagioclase-bearing olivine websterite; 28, plagioclase-bearing lherzolite.

(a) Net projection of the plagioclase (Pl)-olivine (Ol)-orthopyroxene (Opx)-clinopyroxene (Cpx) tetrahedron. (b) Pl-Ol-Opx-Cpx tetrahedron showing the position of the 50% plagioclase section and the section A-B-C. (c) Section through the tetrahedron at 50% plagioclase. (d) Section A-B-C through the tetrahedron.

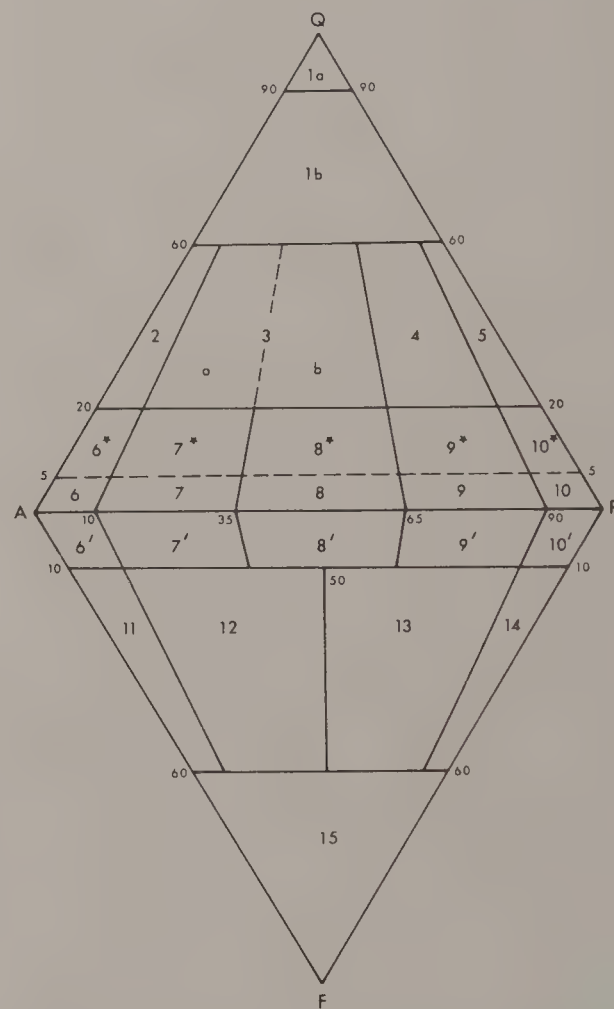


FIGURE 2. The Q(quartz)-A(alkali-feldspar)-P(plagioclase)-F(feldspathoid) diagram for plutonic rocks where M < 90%.

Root names: 1a, quartzolite(silexite); 1b, quartz-rich granitoids; 2, alkali-feldspar granite; 3, granite; 4, granodiorite; 5, tonalite; 6, alkali-feldspar syenite; 7, syenite; 8, monzonite; 9, monzodiorite-monzogabbro; 10, diorite-gabbro-anorthosite (for 6*-10* prefix with "quartz-", 6'-10' prefix with "feldspathoid-bearing" or "foid-bearing"); 11, foid or feldspathoid syenite; 12, foid or feldspathoid monzosyenite; 13, foid or feldspathoid monzodiorite-monzogabbro; 14, foid or feldspathoid diorite-gabbro; 15, foidolite or feldspathoidolite.

- Q Quartz plus any other silica polymorph.
 A Alkali feldspar, consisting of orthoclase, microcline, perthite, anorthoclase, albite (An_{00-05}) and sanidine.
 P Plagioclase (An_{05-100}) plus scapolite.
 F Feldspathoids or foids, i.e., leucite and pseudoleucite, nepheline, sodalite, nosean, hauyne, cancrinite, analcime, zeolite, etc.
 M Mafic and related minerals (micas, amphiboles, pyroxenes, olivine, opaque minerals, garnet, melilite, monticellite, accessory minerals, and primary carbonate).

The first-order subdivision is based on the proportion of M group minerals in the total. The cut-off is given as $M=90\%$, and rocks exceeding this value are termed "ultramafolite" or "ultramafic rocks," and given a specific name according to the principal phase(s) involved.

Ultramafolites

Ultramafolites have $M \geq 90\%$, $Q + A + P + F \leq 10\%$ and are composed of olivine (ol), orthopyroxene (opx), clinopyroxene (cpx), hornblende (hbl), biotite (bio), with lesser proportions of garnet, spinel, and opaque phases (Fig. 1).

Peridotites and *pyroxenites* are rocks in which $ol + opx + cpx \geq 90\%$ and $hbl + bio + \&c. \leq 5\%$. Peridotites contain $<40\%$ opx + cpx, pyroxenites contain more. The root names are *dunite*, *orthopyroxenite*, *clinopyroxenite*, *harzburgite*, *wehrlite*, *lherzolite*, *websterite*, *olivine clinopyroxenite*, *olivine orthopyroxenite* and *olivine websterite*. Where one particular pyroxene species is identified, terms like *diopsidolite*, *augitolite*, and *enstatite* (or *enstatitolite*), substitute for pyroxenite. Hornblende, garnet, and spinel should be used as qualifying prefixes (Table 1). Spinel and ore minerals are ubiquitous accessories and do not prefix a root name if they form $<5\%$ of the rock.

Gabbroids

The M values of gabbroids have a nearly complete range, and the $Q + A + P + F$ values

vary considerably, but the bulk (i.e. $>95\%$ of $Q + A + P + F$) is composed of P, as plagioclase An_{50-100} . Root names are defined on the basis of plag, opx, cpx, and ol proportions, and are *anorthosite* (or *plagioclasite*), *norite*, *gabbro* and *troctolite*. The gabbroids are further subdivided as leuco- ($M=10-35\%$), no prefix ($M=35-65\%$), and mela- ($M=65-90\%$). Where $M > 90\%$, the rocks are termed plagioclase-bearing ultramafolites.

Main QAP triangle

The main QAP triangle covers all rocks consisting of QAPM group minerals, where $M < 90\%$ (Fig. 2). To place a rock in this scheme, $Q + A + P = 100$, and M group minerals are ignored initially. The root names are alkali-feldspar granite, *granite*, *granodiorite*, *tonalite*, quartz-alkali-feldspar syenite, quartz syenite, quartz monzonite, quartz monzodiorite, quartz monzogabbro, quartz diorite, quartz gabbro, alkali-feldspar syenite, *syenite*, *monzonite*, *monzodiorite*, *monzogabbro*, *diorite*, and *gabbro*. The distinction between monzodiorite-monzogabbro and diorite-gabbro is made using plagioclase composition and the nature of the mafic phases. These two pairs occupy the same areas on the QAP triangle. Rocks with $Q/Q + A + P > 60\%$ are rare and problematic; when $Q/Q + A + P$ is 60–90%, they are termed quartz-rich *granitoids*, and when $Q/Q + A + P > 90\%$ they are termed *quartzolite* or *silexite*.

P does not permit discrimination between the various plagioclase species (see gabbro and diorite), so in the granitoid fields other names remain valid, e.g., *trondhjemite* for an oligoclase tonalite or granodiorite. The same holds true for different alkali feldspars grouped as A; thus *plagiogranite* remains a valid name for alkali-feldspar granites dominated by albite.

Foidal rocks: the APF triangle

The silica-undersaturated foidal rocks have historically presented the biggest problems in relation to classification and nomenclature (Fig. 2). Grouped under M and F are very large and diverse collections of minerals, and the use of

TABLE 1. Terminology for some ultramafolites

Root name	% garnet, spinel, etc.			
	$\leq 5\%$	$>5 - \leq 50\%$	$>50 - \leq 95\%$	$>95\%$
DUNITE + garnet	garnet-bearing dunite	garnet dunite	olivine garnetite	garnetite
DUNITE + chromite	—	chromite dunite	olivine chromitite	chromitite

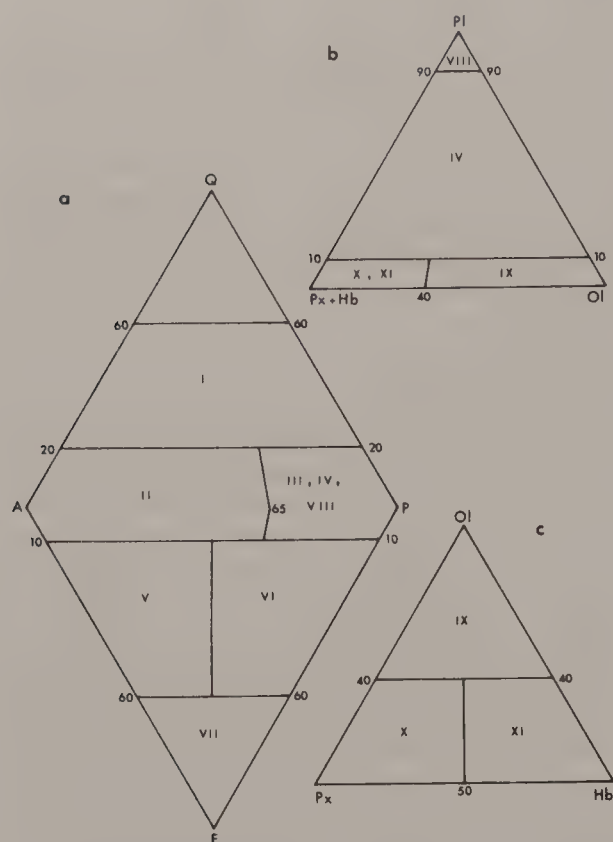


FIGURE 3. Preliminary classification of plutonic rocks for field use: Q, quartz; A, alkali-feldspar; P, plagioclase; F, feldspathoid; Px, pyroxene; Hb, hornblende; Ol, olivine. (a) General classification. (b) Ultramafic, gabbroic, and anorthositic rocks. (c) Ultramafic rocks.

Names: I, granitoids; II, syenitoids; III, dioritoids; IV, gabbroids; V, foid or feldspathoid syenitoids; VI, foid or feldspathoid gabbroids and dioritoids; VII, foidolites or feldspathoidolites; VIII, anorthositic rocks; IX, peridotites; X, pyroxenites; XI, hornblendites. II–V qualified by “foiid-bearing” if feldspathoids present; IX–XI, ultramafic rocks or ultrafolites.

locality-derived names for the vast range of variations is endemic (see Sørensen, 1974).

When $F/A + P + F$ is $<5\%$ the root names gabbro, diorite, monzogabbro, monzodiorite, monzonite, syenite, and alkali-feldspar syenite are used with the prefix “feldspathoid-bearing” or “foiid-bearing.” When $F/A + P + F$ lies between 10 and 60%, these same root names are prefixed by “feldspathoid” or “foiid.” For rocks where $F/A + P + F > 60\%$ the term “feldspathoidolite” or “foiidolite” is used. Greater specificity is achieved when the foid(s) are identified, so that essexite is a foid monzodiorite–foiid monzogabbro or specifically a nepheline monzodiorite–nepheline monzogabbro; likewise, teschenite is a foid gabbro–specifically analcime gabbro.

Charnockite suite

Rocks of the charnockite suite plot in the QAP triangle but are distinguished from the

main group by the presence of orthopyroxene in the mode, where it may be the predominant mafic mineral. The orthopyroxene is usually hypersthene but may include enstatite or bronzite. Clots of quartz + fayalite or quartz + fayalite + magnetite may have formed after Fe-rich orthopyroxene. Root names in general terms are based on the root names used in the standard QAP triangle with the prefix “hypersthene,” but the special terms *charnockite*, *farsundite*, *opdalite*, *enderbite*, *mangerite* and *jotunite* are entrenched in the literature and are retained as alternatives (see **Charnockite suite**).

Hypabyssal rocks

Most holocrystalline hypabyssal rocks have the same modal mineralogy as the coarse-grained, phaneritic plutonic varieties considered above. The IUGS recommendations suggest using the root names defined above with the prefix “micro-” where the grain size in the groundmass is less than 2 mm. The exception is in the rocks of basaltic composition where the root name *dolerite* is retained as an alternative to *microgabbro*. Where glass is present in the rock, the IUGS recommendations for volcanic rocks apply (see **Volcanic rocks**).

Field terms

Streckeisen (1976) recommends a preliminary system of nomenclature specifically for field use. The general names are based on readily identified and roughly quantifiable mineral species, and can be prefixed by terms like “leuco-,” “mela-,” and “micro-.” Many of these terms have the termination “-oid,” or consist of grouped common names from several QAPF fields combined.

ADRIAN F. PARK

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Cross-references: *Acid(ic) igneous rocks*; *Alkaline rocks—undersaturated*; *Basic igneous rocks*; *Igneous rocks—classification and nomenclature*;

Intermediate igneous rocks; Peralkaline igneous rocks; Plutonic rocks; Ultrabasic igneous rocks; Ultramafic rocks.

PNEUMATOLYSIS

The process of *pneumatolysis* involves the transfer of volatile materials from a crystallizing magma to the enclosing wall rocks by means of a gaseous (not liquid) phase evolved from the crystallizing magma. The fact that many fluids escaping from magmas are probably supercritical makes the distinction between liquid and gas somewhat superfluous. Contact metamorphic aureoles containing components of a supposedly volatile nature (B, F, Cl, S, As, Sb), or those transported as volatile compounds (Fe or Sn halides; SiF_4), are frequently referred to as "pneumatolytic." *Autopneumatolysis* involves the crystallization of minerals, or the alteration of a rock, by gaseous emanations originating in the magma or rock itself.

D. R. WONES

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- Cross-references: *Aureole; Contact metasomatism; Granites and mineralization; Greisen; Hydrothermal processes; Metasomatism; Pyrometasomatism; Skarn; Volatiles.*

POLYMETAMORPHISM

Many metamorphic rocks are the result of more than one phase of metamorphism. In fact, most regionally metamorphosed rocks in orogenic belts show the effects of repeated phases of metamorphism (Fig. 1). Despite this, most textbooks give little attention to this aspect of metamorphism and describe only cases of one single metamorphic episode. However the polyphase nature of regional metamorphism is clearly revealed in the textures of the rocks (Fig. 2) and textural relationships can be used to establish both successive mineral growths and the conditions of growth (static or dynamic) (Table 1). Within one orogenic episode this may be due to recrystallization in several separate phases during a single rise and fall in temperature, or it may be due to several separate phases of metamorphism (Fig. 3). These two cases are difficult to distinguish, particularly in the absence of isotopic data.

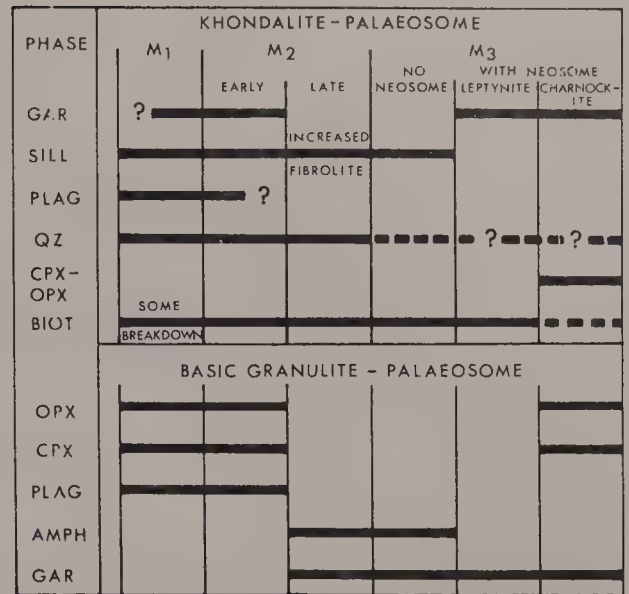


FIGURE 1. Metamorphic mineral growths during three phases of metamorphism (M_1 , M_2 , M_3) in khondalite (garnet-sillimanite gneiss) and pyroxene granulite; Angul, Orissa, India. (From Park and Dash, 1984.)

Such polyphase metamorphism associated with a single orogenic cycle is regarded as polymetamorphism by some authors (Spry, 1969; Turner, 1981), but others confine the term to the superimposition of a separate later episode of metamorphism on an earlier one,

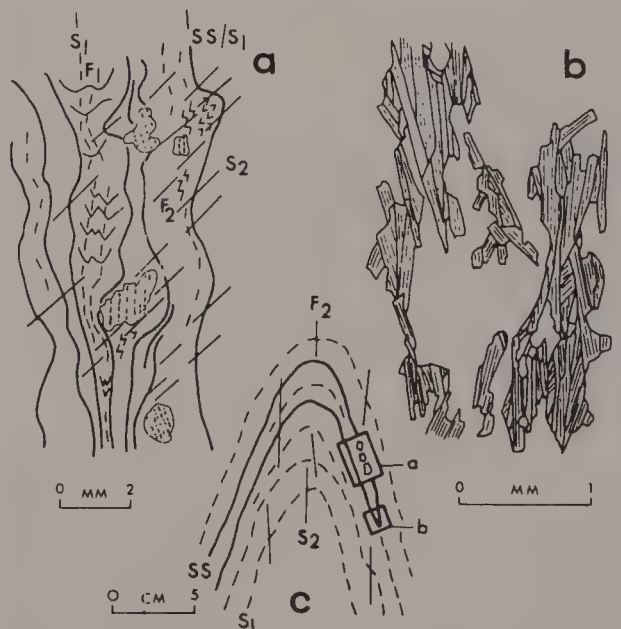


FIGURE 2. Polyphase nature of regional metamorphism in K-feldspar-sillimanite-garnet-biotite gneiss shown by garnets preserving weak planar quartz internal fabric (a), biotite fabric showing the hinge zone of an F_1 fold (b), and S_1 schistosity folded by F_2 fold with the development of a second (S_2) schistosity (c); Tampere, Finland. (From Campbell, 1980.)

TABLE 1. Successively developed metamorphic mineral assemblages in ultramafic rocks, eastern Finland

Igneous protolith		Metamorphic mineral assemblage		
Rock	Minerals	Pre-D ₂	D ₂ dynamic	D ₂ static
Dunite	Olivine, chromite, magnetite, pyrrhotite	Chrysotile, magnetite	Talc-chrysotile schist	Anthophyllite rosettes
Wehrlite	Olivine, diopside, chromite, magnetite	Chlorite, talc, chrysotile	Talc-chlorite schist	Anthophyllite serpentinite
Hornblende, Hornblende?	Clinopyroxene?, hornblende, ilmenite, zircon	Foliar hornblende	Chlorite schist with ilmenite and zircon	Hornblende, phlogopite, chlorite, zircon, ilmenite rock

Source: after Park (1983).

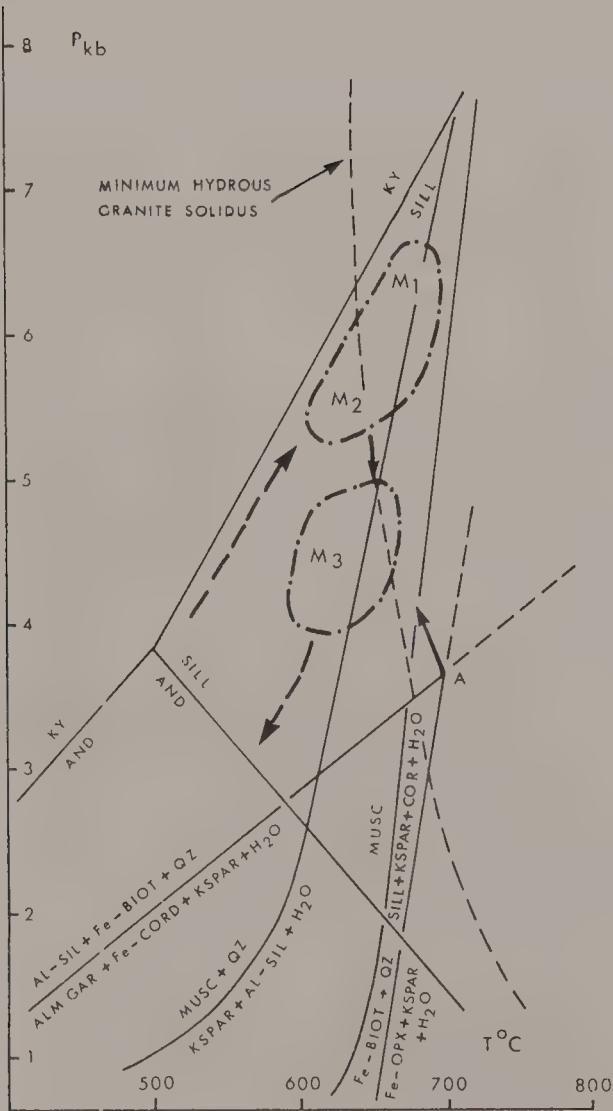


FIGURE 3. Changing *P-T* conditions during the development of the metamorphic mineral growths illustrated in Fig. 1. (From Park and Dash, 1984.)

with no genetic connection between the two (Read, 1949; Hsu, 1955). In this latter category, the effect of two different types of metamorphism is most easily distinguishable, e.g., thermal or cataclastic metamorphism affecting regional metamorphic rocks. However, it is also common for regional metamorphic rocks, particularly gneisses, to be the products of reworking of older regional metamorphic rocks.

Thermal overprints on regional metamorphic rocks

Commonly, high-level igneous plutons are intruded into areas of regional metamorphism. High-grade hornfels near the contact will generally be so completely recrystallized that earlier metamorphic effects are completely obliterated. However, the production of randomly-oriented porphyroblasts of high-temperature minerals is typical of the outer zones of contact metamorphic aureoles, and in these areas, chiastolite slates, cordierite schists, etc., may still preserve the mineral fabric of a regional metamorphism while giving evidence of the thermal effects by the growth of large porphyroblasts. Farther from the pluton, “knots” of indeterminate origin are a feature of low-grade regional metamorphic slates, being aggregates of iron oxides and other minerals induced by the contact metamorphism. Schists and phyllites may also, nearer the pluton, become bleached by the recrystallization of iron oxides and dark micas into randomly oriented porphyroblasts while still retaining the strong banding and even the fabric, in minerals such as muscovite, of their regional metamorphic precursor.

Thermal overprints on cataclastic rocks

Many cataclastic rocks recrystallize statically after the shearing movements have ended. This is termed *neomineralization* and the resultant rocks have been termed *protomylonite gneiss* and *blastomylonite* (Higgins, 1971). The cataclastic fluxion texture is still evident, but the grains are unstrained and have recrystallized to give, normally, a highly interlocked sutured texture. The microscopic evidence for cataclasis has thus been destroyed by a type of annealing process, but this effect should not really be regarded as polymetamorphism. It becomes very difficult, however, to draw a distinction between this effect and the subsequent recrystallization of shear zones or mylonite bands in regional metamorphic areas. An initial phase of strong deformation frequently produces shear zones or mylonites and subsequent metamorphic episodes cause the preferential growth of porphyroblasts in the finely deformed rocks. These overgrowths, typically of biotite or hornblende, may attain sizes of several centimeters without any recrystallization of comparable age being apparent in neighboring unsheared rocks. This is probably due to the extremely fine grain size of the cataclastic rocks, facilitating recrystallization, and also to the residual strain energy available in the deformed grains that is not available in the surrounding rocks. Shear zones also act as natural pathways for the passage of volatiles and this again would encourage crystal growth in these limited zones. Addition of such ions as K and Rb has also been demonstrated in some cases.

Cataclastic as well as regional metamorphic rocks may also be intruded by plutons and thus be subjected to thermal metamorphism. It will again be difficult or impossible to recognize the cataclastic effects in high-grade hornfels, although strong fluxion texture with large frequent augen may be preserved by the thermal recrystallization, but mylonites may only be recognized by their very strong banding. Where different sedimentary rock types are sheared together in a mylonite zone, so that contrasting bands of different chemistry are found (e.g., limestone and quartzite), the subsequent thermal metamorphism may produce new minerals (e.g., calc-silicates) that are not found in the aureole where the rocks were not previously mylonitized.

Cataclastic metamorphism of regional metamorphic rocks

Cataclasis can affect any rock type, but the major belts of extreme deformation where mylonites and cataclasites are principally developed almost invariably occur in basement

rocks at the margins of orogenic belts. Thus, mylonites are best and most typically developed in regional metamorphic gneisses or granulites; some of the most famous examples occur in more recent orogenic belts, e.g., the Moine thrust zone of NW Scotland, where gneisses of the Precambrian Lewisian complex were extensively deformed and mylonitized in the Caledonian orogeny. Belts of mylonite and cataclasite are a notable feature of Precambrian orogenic belts, many of them vertical in attitude and many kilometers wide (Davidson, 1984). It is principally in protomylonites (Higgins, 1971) with a high proportion of augen of pre-existing rock that the earlier regional metamorphic effect can be distinguished. Once the rocks have become very strongly deformed to mylonites or ultramylonites, there is little evidence of the polymetamorphic nature of the rock type.

Polymetamorphism by repeated regional metamorphism

The superimposition of a different type of metamorphism upon a previously metamorphosed rock will be likely to result in textures or mineralogy that will enable the presence of the two episodes to be distinguished. In the case of regional metamorphism, this is far less likely. As a natural result of the thermodynamics of the system, progressive metamorphic reactions proceed to completion comparatively quickly and mineral stabilities are generally quickly attained under conditions of increasing temperature and pressure. Therefore if a low-grade metamorphic rock is subjected to a higher grade of metamorphism in a subsequent orogeny, it will tend to be completely recrystallized. Only where higher-grade rocks are subsequently subjected to lower grades of metamorphism are relict textures and mineralogy commonly preserved. In these cases there is a very strong control over recrystallization by the combined effects of water availability and concomitant strain.

One area where two high-temperature metamorphic events occur without the later one destroying the effects of the earlier is in Connemara, Eire. Here the normal sequence of multiple phases of deformation and metamorphism reaches its climax during the second phase of deformation at kyanite grade in the pelitic rocks, as it does over large areas of Scotland. However, superimposed onto this during the third deformational episode (which elsewhere usually occurs during a waning retrogressive phase of metamorphism), there occurs the growth of andalusite and cordierite, suggesting that a phase of Buchan-type metamorphism (high T -moderate P) had succeeded

the earlier Barrovian-type metamorphism (high T -high P) (Yardley, 1976). This has been ascribed to the effects of the emplacement of a regional complex of synorogenic basic plutons causing a local increase in the geothermal gradient.

There is increasing evidence of the polyphase nature of most gneiss terranes, particularly the reworked nature of parts of many Proterozoic orogenic belts (Myers, 1984). This is partly due to the occurrence of a widespread late Archaean phase of granulite facies metamorphism 3,200–2,800 Ma ago that resulted in recognizable relics of high-grade rocks of this age being preserved in reworked terrane of younger ages. The granulite facies mineralogy can readily be recognized in retrogressed rocks, but it is also found that the granulite facies rocks acquired a distinctive trace element geochemistry, involving depletion in K, Rb, U, Th, and Cs (Tarney et al., 1972), which also acts as a marker.

Crustal reworking of the earliest Archaean rocks has been studied in detail in W Greenland (McGregor, 1973). In this area, rocks have been extensively reworked several times and older gneisses can be seen to pass gradually by increasing amounts of deformation and metamorphism into the newer gneiss complexes, which include the older gneisses and deformed relics of intervening phases of intrusive dolerites and granites and unconformably overlying sedimentary rocks. In addition, isotopic studies have shown that the original gneiss complex was successively added to by plutonism and that some of the rock assemblages interpreted from field evidence as being "reworked" are new crustal additions of mantle-derived material (Moorbath, 1977). The region shows a long history of Archaean metamorphic and igneous events as well as later reworking during Proterozoic times.

A. E. WRIGHT

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Cross-references: *Aureole*; *Cataclastic rocks*; *Contact (thermal) metamorphism*; *Migmatite—structural relationships*; *Polyphase metamorphic mineral growth*; *Regional metamorphism*; *Retrograde metamorphism*.

POLYPHASE METAMORPHIC MINERAL GROWTH

During orogenesis, regionally metamorphosed terranes are normally held at elevated temperatures and high confining pressures for an extended period. Sporadic and probably relatively short-lived phases of penetrative deformation, involving directed pressures, are, therefore, interspersed by possibly longer periods of quiescence during which the main external physical factors influencing the crystallization of metamorphic minerals are temperature and confining pressure. Deformation itself breaks down existing fabrics, sedimentary, igneous, or metamorphic and reduces grain size. By doing so it must act as a catalyst both for reaction and recrystallization; total grain surface area is considerably increased during deformation and abundant high-energy sites are provided not only by this increase but by internal strain induced in crystals. Periods of static recrystallization between phases of deformation are responsible for the coarsening of metamorphic rocks as they strive for an

equilibrium state, eliminating by recrystallization the strain inherent in deformed crystal lattices, and by grain growth, minimizing strain at crystal boundaries. Thus, what is normally the most conspicuous feature of a regionally metamorphosed rock—its planar and/or linear fabric—was induced during relatively short-lived phases of deformation, whereas its mineral assemblage is the product of static periods during which temperature and confining pressure were the main external factors that controlled progress towards an equilibrium state.

Much modern work on metamorphic rocks has concentrated on mineral assemblages, and used experimental work to discover the physical parameters that control critical metamorphic reactions. Hence, it is possible to provide estimates of the P – T conditions under which many regionally metamorphosed rocks crystallized. Such work reveals the P – T conditions at the peak of the metamorphism, i.e., at one stage in a complex history. Metamorphic textural studies, by reaching back beyond this single stage, can supplement work on facies and provide information about the early metamorphic and structural history. Pioneering modern work on regional metamorphic textures was carried out by Rast (1958) and Zwart (1960) and their conclusions and those of other workers are summarized by Spry (1969). These conclusions show that fieldwork is profitably continued in the laboratory with the use of oriented specimens, so that structures identified and measured in the field can be investigated in thin section and their relationship to shape, distribution, and orientation of the mineral constituents can be determined. Pelitic and calcareous metasediments and metabasaltic rocks have proved particularly suitable for this type of study, largely because their chemical composition allows the development of different phases (often as porphyroblasts) as metamorphism proceeds, and because they contain minerals whose crystallographic orientation can be easily seen because of conspicuous cleavage or shape. Rocks such as quartzites have yielded interesting fabrics, but their preferred orientation can often only be detected laboriously on the universal stage.

Phases of deformation (D_1 , D_2 , etc.), being relatively short-lived and, over limited areas probably contemporaneous events form valuable time-markers in the metamorphic history of an area. The relationship between phases of post-tectonic metamorphism (MP) during which there was fabric-coarsening or porphyroblast growth, and of syntectonic metamorphism (MS) during which dimensionally-oriented mineral fabrics were formed can be deduced from

	MS1	MP1	MS2	MP2	MS3	MP3	MS4	MP4	MS5	MP5
a										
QZ		— ? —								
MUS		— ? —								
BIO		— ? —								
GAR		— ?								
SILL										
PLAG/KSP		— ? —								
b										
QZ		— ? —						— ? —		
BIO		— ? —						— ? —		
PLAG/KSP		— ? —						— ? —		
GAR		— ?	— ?							
SILL										
MUS								— ?		
c										
QZ		— ? —								
BIO		— ? —								
PLAG/KSP		— ? —								
CORD	— ?		— ?							
GAR	— ?		— ?							
SILL										
MUS										

FIGURE 1. Syntectonic (MS) and posttectonic (MP) mineral growth sequences during five deformational phases of the Svecokarelian orogenic episode with MS2–MP2 the climactic metamorphic event: (a) in muscovite–sillimanite zone; (b) in K-feldspar–sillimanite zone; (c) in K-feldspar–cordierite zone; Tampere district, Finland. (From Campbell, 1980.)

metamorphic textures (Fig. 1).

When describing textures, Si (internal) is used to refer to fabrics poikiloblastically enclosed within a porphyroblast, while Se (external) is used for groundmass fabrics. The relationship between Si and Se is of great importance in textural interpretation. Many mineral species are able to grow as porphyroblasts during periods of static crystallization, garnet, kyanite, staurolite, andalusite, amphibole, and plagioclase being the most common. By contrast, garnet is the only common syntectonic porphyroblast.

Post-tectonic mineral growth

At low metamorphic grades, pelitic rocks usually exhibit a slaty cleavage that consists of the alignment of flaky minerals with (001) lying in or near the XY plane of the finite-strain ellipsoid (Ramsay, 1967). Commonly, very small elongate quartz grains of clastic origin show a strong preferred orientation in that plane. In coarse pelitic rocks at higher grades of metamorphism, the preservation of this fine-grained oriented quartz fabric within early-formed porphyroblasts (Fig. 2a) may constitute the only, but none-the-less important evidence that high-grade rocks progressed fairly gradually through the lower grades during metamorphism. Traces of bedding, obliterated in the groundmass may be detected within early-formed porphyroblasts (Fig. 2b) occurring as graphitic

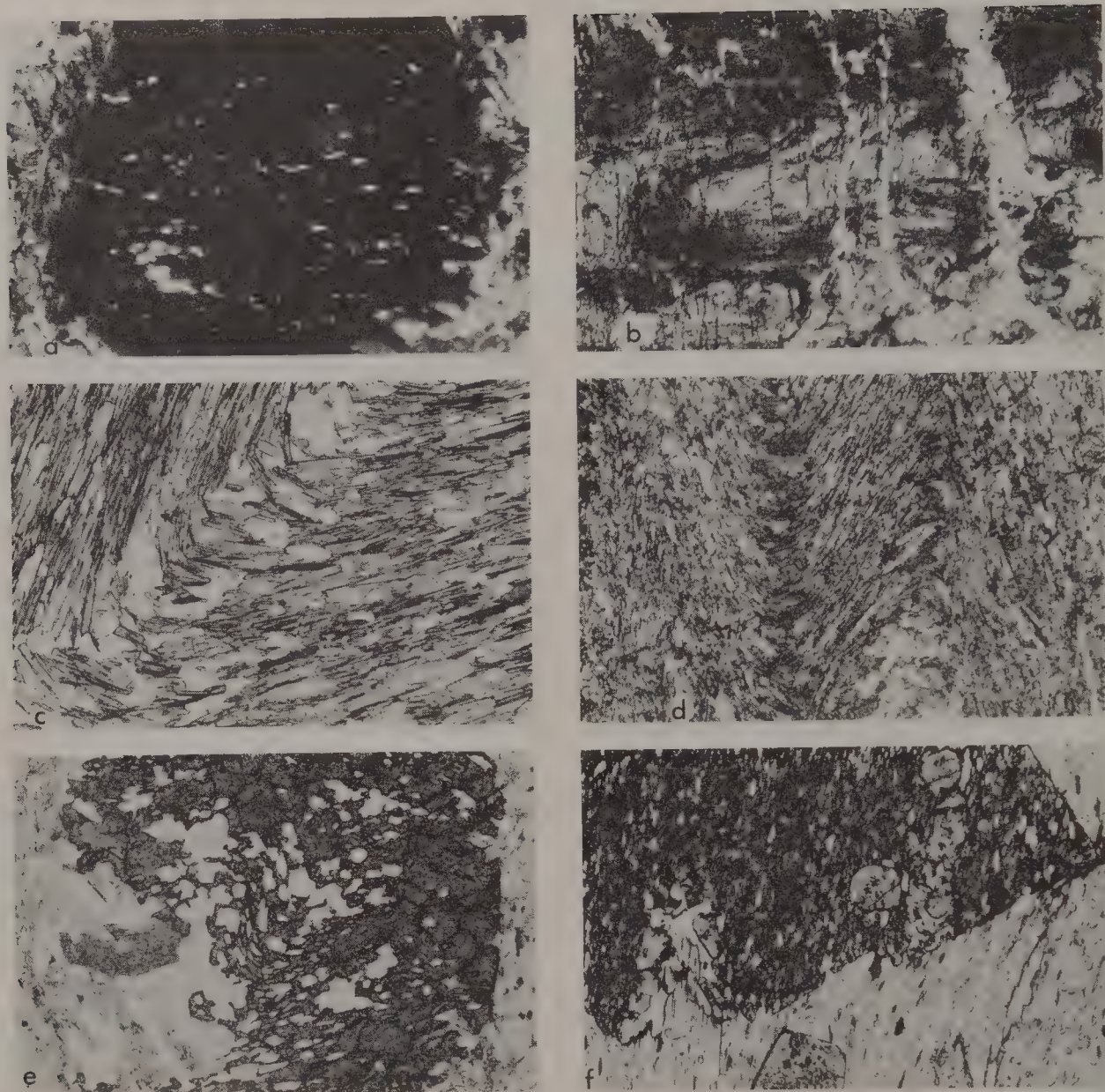


FIGURE 2. (a) Post-tectonic garnet porphyroblast overgrowing a dimensionally-oriented quartz fabric; XN, $\times 9$. (b) Post-tectonic hornblende porphyroblast overgrowing folded graphite trails (bedding); PPL, $\times 9$. (c) Buckled planar fabric defined by mica; quartz concentrated in the hinge zone of the fold; no individual mica flake is bent; PPL, $\times 9$. (d) Buckled linear fabric defined by amphibole needles; no individual amphibole prism is bent; XN, $\times 9$. (e) Post-tectonic garnet porphyroblast that has overgrown a folded quartz fabric; XN, $\times 9$. (f) Post-tectonic poikiloblastic porphyroblast of staurolite that has overgrown a planar quartz-mica fabric and garnet porphyroblasts; PPL, $\times 7$.

laminae in carbonaceous metasediments, as grain-size variation in clastic sediments, or as compositional heterogeneity brought out by laminae containing, for example, more or less epidote, opaque ore, or apatite. The remnants of cleavage, schistosity, or folded bedding preserved as Si within porphyroblasts indicate that the latter are post-tectonic. (Fig. 2a,b,e,f).

Although regionally metamorphosed rocks show many indications of deformation, such as tectonically induced planar and linear structures

and minor crenulations, in most cases individual crystals show little sign of internal strain. It is extremely unlikely that strained micas and quartz grains that display undulose extinction or calcite grains with bent twin lamellae have suffered this deformation during or before the peak of regional metamorphism. Strain induced at this early stage was almost certainly completely resolved by crystallization and grain growth; the strained material now recognizable probably underwent deformation at a late stage

when the rocks were much more brittle and at low temperatures so that annealing processes have not operated. With regard to structures formed at early stages of orogenesis, it is a remarkable feature that detailed examination of tightly folded schistosity surfaces defined by oriented mica flakes comprising a *lepidoblastic* texture reveal that no individual flake is deformed, although collectively they define a folded surface (Fig. 2c). Similarly, in hornblende schists that have undergone polyphase deformation, the strong *nematoblastic* fabric formed parallel to the stretching direction of an early phase is commonly buckled during subsequent folding (Fig. 2d). Although the fabric is strongly deformed, it is commonly observed that no individual amphibole crystal is bent, though it is difficult to conceive that at the time of deformation individual crystals were not strained. Clearly, recrystallization has healed the deformed crystals of mica or amphibole while allowing them to retain a high degree of preferred orientation. Thus, the nucleation that initiated the new crystallization must have been to some extent subject to tectonic control and there is no reason to consider that phases other than amphibole and mica were not similarly subject. Crystallization of this type is referred to as *mimetic* and has been compared to the stage of primary crystallization recognized in hot-worked metals (Spry, 1969).

Similar control by tectonic surfaces is shown by the frond-shaped arrangement of amphibole crystals (hornblendegarbenschiefer) that sometimes lie within micaceous schistosity surfaces. These amphibole crystals do not spear through the surfaces but lie more or less randomly within them. Grain growth of the mimetic fabric may retain its preferred orientation, but commonly porphyroblasts consisting of new or already established phases, which formed during the post-tectonic stage, grow with random orientation across tectonic surfaces and lines. It is these that are likely to contain relict inclusions of preferred orientation (e.g., Fig. 2a), but even if relict inclusions are not present, post-tectonic porphyroblasts may be inferred where the porphyroblast abruptly truncates the oriented groundmass fabric. Porphyroblasts around which foliation is deflected are likely to be pre-tectonic. Grain growth commonly results in a reduced degree of preferred orientation and is a feature of metabasaltic rocks among others. Metabasaltic rocks that have been thoroughly deformed show a strong degree of preferred orientation of amphibole crystals with the *c* crystallographic axis being generally parallel to the stretching (*X*) direction of the finite-strain ellipsoid. In such hornblende schists, other minerals (epidote, sphene, and ore) often show

dimensional orientation related to that of amphibole. Commonly epidote, sphene, or ore retain an orientation in rocks in which amphibole orientation has ceased (by grain growth) to be conspicuous; this suggests that different phases possess a different capacity to retain preferred orientation.

Porphyroblasts that contain trails of inclusions having a well-marked dimensional orientation have grown over a strong planar or linear fabric—a cleavage, schistosity or, in rare cases, bedding. Such inclusions may be the only fine-grained strongly-aligned grains that survive, having been protected from the crystallization and grain growth that has coarsened the groundmass material and removed its strong dimensional orientation. Rarely it is possible to use the included material to trace progressive changes in the groundmass. In such cases the central part of the porphyroblast contains small strongly aligned inclusions, while successive peripheral zones show an increasing grain size and/or a decreasing degree of dimensional orientation (cf., Bell and Johnson, 1989).

It is sometimes argued that, where porphyroblasts (say, of garnet) appear to have overgrown a fine slate or phyllite fabric, the minute quartz inclusions are the remains of much larger crystals, and that their small size compared to peripheral inclusions or to groundmass material simply reflects that, at an early stage, garnet absorbed, or otherwise disposed of, silica much more efficiently than at a later stage. In many cases, however, very small inclusions, each of differing crystallographic orientation, are much smaller than groundmass crystals, even when their size is conceived to be increased until they are impinging on neighboring inclusions. Further, it is more reasonable to conclude that strongly-oriented included grains, very similar to those found in slates and phyllites, are the remains of slate and phyllite fabrics only slightly modified during their incorporation into porphyroblasts.

Whereas the earliest planar tectonic structure to form is a slaty cleavage or flow schistosity, subsequent phases of deformation are accompanied by crenulation cleavage. Where reworking is very thorough, the crenulation cleavage surfaces may be very closely spaced, simulating in the field a slaty cleavage, although its secondary nature is readily detected in thin section. Consequently, where inclusions indicate that porphyroblasts have overgrown crenulation cleavage (Fig. 2e) it may be inferred that the porphyroblast is post-tectonic with respect to a second or subsequent phase of deformation. A similar inference may be drawn where Si patterns show growth over a folded cleavage fabric.

Syntectonic mineral growth

It is important to distinguish between porphyroblasts that, having grown over a folded fabric, are post-tectonic, and those that are syntectonic, having grown during deformation. Many porphyroblasts grow over a "pair" of microfolds and thus have symmetrically arranged curving S- or Z-shaped trails of inclusions that are extremely difficult to distinguish from S- and Z-shaped inclusions incorporated into a crystal rotating during growth. In such cases, syntectonic crystals can be distinguished from post-tectonic ones by studying thick serial sections through individual crystals on a universal stage (Powell and Treagus, 1970). In this way it can be discovered whether the inclusions have a spindle-shaped arrangement (indicative of syntectonic growth) or whether approximately cylindroidal folds can be traced through the crystal (indicative of post-tectonic growth). Powell and Treagus also point out that syntectonic crystals can be recognized by $\parallel$ -shaped $\cup$ -shaped and $\cap$ -shaped inclusion trails in sections cut parallel to or at low angles to the axis of rotation. Porphyroblasts such as the snowball garnet in Fig. 3a must also result from rotation. Curving inclusion trails that would imply rotation much in excess of 180° cannot readily be explained by the overgrowth of a folded fabric. Ramsay

(1962) has pointed out that progressively increasing flattening deformation of the matrix fabric during porphyroblast growth will also result in curving inclusion trails. A remarkable example of syntectonic growth has been described by Zwart (1963), who demonstrated porphyroblast growth during the progressive buckling of a foliated matrix.

In the simplest case, the sense of rotation of each syntectonic porphyroblast is the same on the same fold limb, curving Si trails being continuous with Se fabric. Often these relationships are obscured by renewed rotation of a syntectonic crystal after its growth was completed. This has the effect of disrupting the continuity of Si with Se. Post-tectonic grain growth of Se may also make this continuity less obvious.

Pretectonic mineral growth

Porphyroblasts that have grown later than the formation of a planar structure do not appear to push aside the pre-existing fabric during growth. Force of crystallization in this sense is absent from regionally metamorphosed rocks, as may be reasonably inferred by the presence of the most delicate Si bedding lamination and cleavage fabrics (e.g., Fig. 2a,b). It is concluded that where planes of cleavage or schistosity wrap around porphyroblasts, such crystals are pre-tectonic and, acting as relatively rigid nuclei set possibly in a fairly ductile pelitic matrix, have caused the localized deflection of the new cleavage. This generally corresponds to the way a pebble deflects planes of cleavage in a deformed sediment, or a crystal deflects eutaxitic texture in a welded ash flow. Such deflection in a metamorphic rock may accompany the formation of augen of which the pre-tectonic porphyroblast forms the center, the low-pressure corners being filled with readily soluble minerals such as calcite or quartz.

Porphyroblasts are often associated with deflected matrix schistosity in quite a different manner. In such cases a crystal has grown over a foliation that has subsequently been deformed either by the flattening of the Se ductile fabric against the rigid porphyroblast or by rotation of the latter after its growth is complete. Such deformation results in disruption and deflection of the Se fabric in the immediate vicinity and the movements, if intense, can produce an augen effect with porphyroblasts completely detached from, and lying within, deflected foliation. The geometry may be used as a criterion indicating sense of shear. In this case, however, Si and Se are of the same age, but Si trails lie at a discernible angle to the Se while the peripheral zones may show signs of

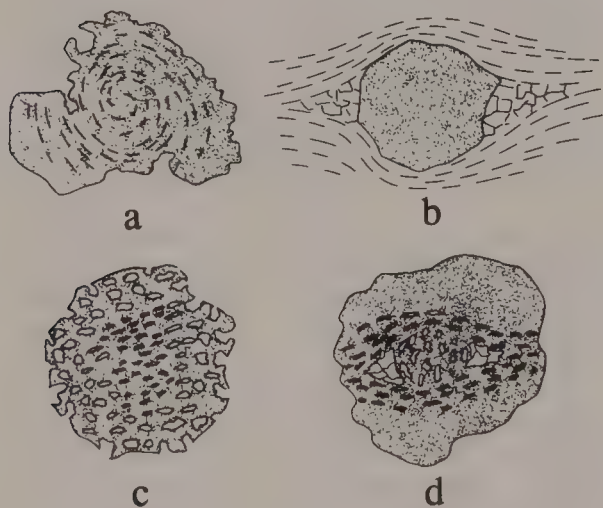


FIGURE 3. (a) Snowball garnet (syntectonic) with pattern of inclusions suggesting growth during rotation of up to 540° . (b) Pretectonic garnet porphyroblast; schistosity in the groundmass is deflected around the garnet to form an augen structure; quartz has grown in the low-pressure areas. (c) Progressive post-tectonic garnet growth; early stages grew over a dimensionally-oriented fabric; later stages trace the changes in the groundmass to larger, equant grains. (d) Pre- and post-tectonic garnet; the early stage was augened (see (b)); subsequent garnet growth has overgrown the deflected foliation.

retrogression, e.g., garnet rims converted to chlorite. The presence of numerous porphyroblasts each containing trails of inclusions, the orientation of which may vary from one crystal to the next, indicates that at an early stage the porphyroblasts overgrew a planar foliation but that the extent of subsequent mechanical rotation has varied from crystal to crystal.

Assessment

Polyphase growth of individual metamorphic minerals is common. Thus, the central zone of a garnet may contain minute strongly-aligned quartz inclusions produced when the early stage of the crystal overgrew a slate fabric. The core zone may be succeeded abruptly (Fig. 3c) by a peripheral zone in which large equidimensional grains of quartz are included, perhaps of a size consistent with the present grain size of the matrix. Such a crystal preserving two stages in the evolution of the groundmass has itself clearly grown at two different stages: the first was very early in the metamorphic history and the second when grain growth in the matrix had virtually ceased. Two such stages may show changes in composition (Atherton, 1968). In other cases, syntectonic crystals may have peripheries having large randomly-disposed inclusions, while porphyroblasts augened at an early stage may have peripheral zones that have overgrown the augen structure (Fig. 3d).

The relative age of episodes of mineral growth can also be detected by textural studies. Thus (1) Fig. 2f shows garnets that grew before the large plate of staurolite that encloses them, while at the same time replacing the porphyroblasts of biotite, and (2) Chinner (1973) records chloritoid enveloped in garnet and so predating it.

The concept that metamorphic mineral assemblages accurately reflect the metamorphic facies conditions presupposes that the rocks attained equilibrium. The presence of early-formed porphyroblasts containing an early tectonic fabric must suggest that equilibrium is not fully attained in many cases, since these probably formed under lower facies conditions than those eventually attained. Such a suggestion is strengthened where early-formed porphyroblasts are enclosed by later-formed ones and thus protected from reaction with the matrix under higher facies conditions (e.g., Chinner, 1973).

Consequently, the preservation of textural evidence for several phases of mineral growth suggests that the mineral assemblage preserved may be a disequilibrium assemblage, to be treated with caution when drawing conclusions about metamorphic facies. Similarly, although

there is a tendency for high-grade assemblages to survive the relatively slow cooling after the metamorphic peak, the assemblage may be augmented by retrograde phases induced by hydration or cataclasis. The textural relationship of retrograde minerals to the high-grade minerals often clearly shows their secondary nature; these alteration products can be neglected when assessing the original metamorphic facies.

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Cross-references: *Metamorphism—controls; Nucleation of metamorphic minerals; Polymetamorphism; Porphyroblast; Regional metamorphism; Retrograde metamorphism; Snowball garnet.*

PORPHYROBLAST

A *porphyroblast* (porphyry, plus Greek *blastos*, to sprout) is a large crystal that has grown in a metamorphic rock (Fig. 1; Becke, 1903). The term megacryst refers to a large crystal and is not restricted to metamorphic crystals. The terms maculose, knotted, and nodular are generally synonymous with porphyroblastic. The French term is *phénoblaste* (Schieferdecker, 1959). The use of *blastos* as a suffix in “porphyroblast” must be contrasted with its use as a prefix in “blastoporphyratic,” which refers to a relict texture produced by the partial metamorphic modification of a porphyritic texture, the large crystals being relict igneous and not metamorphic. Rarely, and confusingly, pseudophenocryst and pseudoporphyratic have been used as synonyms of porphyroblastic.

The porphyroblast in a metamorphic rock is analogous to the phenocryst in an igneous rock, although the reasons for the large size differ. A phenocryst grows as an isolated, undisturbed crystal in a fluid medium, and generally owes its size to its early nucleation and long period of growth. A porphyroblast grows in a solid medium and must enlarge itself through an impeding matrix that is itself crystallizing at the same time. It does not owe its size to early formation or necessarily to long, slow growth, because many porphyroblasts form rapidly at a late stage of metamorphism.

Porphyroblast formation is best understood in terms of the relative rates of nucleation (let N be the number of new crystals per unit time) and growth (let G be the increase in size per unit time). A small value of N/G means a few large crystals and a large N/G means many small crystals. Porphyroblasts are the result of the association of one mineral having small N/G with one or more minerals having a large N/G . Garnet, feldspar, cordierite, andalusite, staurolite, and chloritoid tend to form porphyroblasts, whereas quartz, carbonates, and micas tend to be matrix-formers.

The tendency for a given mineral to form porphyroblasts has been attributed in the past to a large “force of crystallization” and has been coupled with the tendency to develop crystal

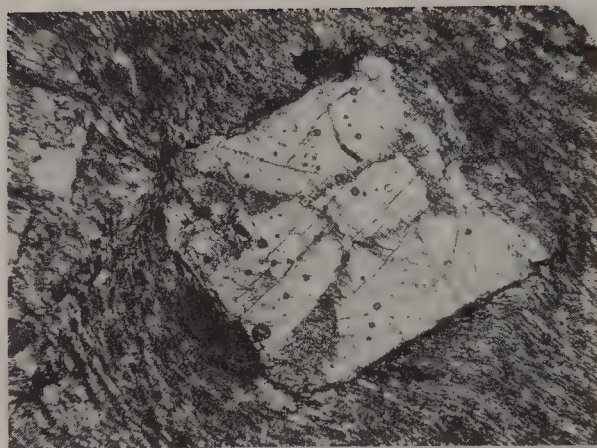


FIGURE 1. Idioblastic porphyroblast of chialstolite in a thermally metamorphosed phyllite; PPL, $\times 19$.

faces, which was attributed to a large “power of crystallization” or “form energy.” The occurrence of large crystals of cordierite that are xenoblastic and of small crystals of spinel that are idioblastic show that the size and shape of a crystal are not directly related. They are related only in that both depend partly on the surface energy of the mineral, and they may be independent because each may be strongly affected by the chemical and/or physical conditions and by the duration of growth. Minerals with a high surface energy tend to be large because they are difficult to nucleate, but difficulty of nucleation is also caused by a small entropy change in crystallization or by the difficulty in forming a nucleus of an atomically complex mineral. Minerals with high surface energy also tend to develop crystal faces, but a slow rate of growth and a high anisotropy of surface energy appear to be more important.

There is controversy over the mutual effects of crystal faces and the enclosing matrix. In general, porphyroblasts grow by extending their boundaries through the surrounding matrix and their rate of growth depends on the ease of attachment of new material and incorporation of inert inclusions (interface control) and by the arrival of required material (diffusion control). Thus, most porphyroblasts grow by replacement but evidence is accumulating to suggest that certain crystals force the adjacent matrix aside (the displacement hypothesis) or grow by replacement plus displacement. Structures in the matrix may be retained as helicitic textures, indicating replacement without distortion.

Most examples of distortion of adjacent S -surfaces are due to deformation after formation of the porphyroblast; some are due to porphyroblast nucleation at pre-existing crumples but a few appear to be due to displacement during growth.

Porphyroblasts are rare in rocks composed of one mineral but isolated large crystals may form in marble, quartzite, or rock salt giving patchy (grumous, clotted, or nodular) texture. This may be the result of either sparse nucleation at a very early stage of recrystallization or abnormal grain-growth (analogous to secondary recrystallization in annealing) at a very late stage of recrystallization.

Porphyroblasts in regional metamorphic rocks may be pre-tectonic (being cracked or otherwise strained, and wrapped around by the foliation), syntectonic (possibly containing rotational structures), or post-tectonic (being cross-cutting superimposed on the foliation, possibly helicitic, and idioblastic).

In the geologic literature of central and northern Europe, and in older literature from the English-speaking world, the nomenclature for porphyroblastic schists, phyllite, and slate uses a German terminology reflecting the type, shape and size of the porphyroblasts. These rocks are commonly metamorphosed tuffs. Thus, *fleckschiefer* has small spots or flecks of indeterminate minerals (cf., spotted slate), *fruchtschiefer* has spots the size of wheat grains, in *garbenschiefer* porphyroblasts are the size of caraway seeds, and *knotenschiefer* has even larger porphyroblasts. These names are qualified by a mineral prefix, e.g., hornblendegarbenschiefer and garnetknotenschiefer.

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Cross-references: *Gneiss*; *Metamorphic mineral growth*; *Nucleation of metamorphic minerals*; *Phenocryst*; *Polyphase metamorphic mineral growth*; *Porphyroblast*; *Schist*.

PORPHYROCLAST

A *porphyroblast* is a relict crystal in a dynamically metamorphosed rock; its derivation appears to be from porphyroclastic texture (Becke, 1903—reference in Spry, 1969) which is equivalent to mortar texture. Porphyroclasts occur as larger crystals set in a finer-grained matrix formed by the crushing, cataclasis, or mylonitization of a rock containing minerals



FIGURE 1. Strained and partially recrystallized lenticular porphyroclasts of quartz in a metasedimentary schist; XN, $\times 1$.

with differing physical properties. Minerals with high strength tend to remain as strained, relict porphyroclasts whereas those that are ductile or plastic or tend to be crushed or recrystallized, form the matrix that flows around them. Porphyroclasts are pre-tectonic crystals (i.e., they have suffered postcrystalline deformation), and hence may be strained (with undulose extinction), contain deformation-lamellae, deformation-twins, deformation-bands, kink-bands, cracks, exsolved phases, bent cleavages or twins, have abnormal optical properties, a lattice and/or dimensional preferred orientation, a ragged, round or elongate shape, and are bounded by pressure-shadows or pressure-fringes (Fig. 1).

The term porphyroclast should not be confused with phenocryst, which refers to a large conspicuous clastic fragment in a poorly sorted sediment.

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Cross-references: *Cataclastic rocks*; *Gneiss*; *Porphyroblast*; *Schist*.

PORPHYRY

A *porphyry* is an igneous rock that contains large crystals (phenocrysts) in a finer-grained groundmass. The name is derived from Greek *porphyra*, which refers to either a small

gastropod or the purplish dye derived from it, hence to the rock (*porfido rosso antico*) from Djebel-Dobhan in Egypt, which has feldspar phenocrysts in a deep red or purplish groundmass (Challinor, 1961). The term porphyritic was apparently first used in a petrographic sense by von Leonhard in 1823.

The term porphyry is generally restricted to rocks with dominantly light-colored (felsic) phenocrysts, most rocks with dominantly dark-colored (femic) phenocrysts being grouped with the lamprophyres. The name of the mineral (or minerals) that forms the phenocrysts may be used as a prefix, e.g., quartz porphyry. One system of terminology contrasts porphyrite (containing feldspar that is predominantly plagioclase, i.e., calc-alkaline) with porphyry (containing feldspar that is predominantly alkaline). The term *porphyroid* has been applied to metamorphic rocks (generally foliated with small augen of quartz or feldspar) that are altered porphyries, porphyritic lavas, or tuffs, or which resemble them.

The term porphyry has a genetic connotation in that it is restricted to hypabyssal rocks: feldspar porphyry is contrasted with its plutonic and volcanic equivalents, porphyritic syenite and porphyritic trachyte.

The contrast in size of phenocrysts and groundmass crystals is due to an abrupt change in cooling rate of the crystallizing magma. The size of any precipitated crystal is a function of N/G and t , where N is the rate of nucleation, and G is the rate of growth, operating for a time t . Many porphyritic rocks have a two-stage crystallization history: an early stage of slow crystallization (presumably due to slow cooling, possibly in a deep magma chamber) followed by a later stage of rapid cooling. The first stage is marked by the free growth of nuclei of minerals that crystallize first, i.e., those of greatest abundance and/or highest melting point. Slow cooling results in a small amount of supersaturation, hence sparse nucleation and slow growth. The second stage is characterized by a high cooling rate, considerable undercooling and high supersaturation of all constituents. This causes abundant nucleation followed by rapid but very short-lived growth to give the fine-grained groundmass. Very rapid chilling, particularly of acidic magmas, may give a glassy groundmass.

Porphyries and mineralization, particularly of Cu and Mo, are commonly associated (see **Granites and mineralization**).

The following are other related terms.

Elvan (*elvanite*)—a Cornish (England) term for a quartz porphyry or a porphyritic microgranite.

Gieseckite-porphyry—a porphyritic rock with

nepheline phenocrysts replaced by scaly sericitic aggregates.

Lenne porphyry—a group name for keratophyres and associated crystal-tuffs, Lenne Valley, W Germany.

Leopardite—a quartz porphyry with quartz phenocrysts in a quartz-orthoclase-albite-mica groundmass; staining by Fe and Mn hydroxides causes a streaked or spotted appearance.

Ortlerite—a hornblende porphyry with a feldspathic groundmass.

Pilandite—a porphyry with abundant anorthoclase as phenocrysts and in the groundmass; the hypabyssal equivalent of hatherlite (Pilandsberg, S Africa).

Porphyrite—a plagioclase (oligoclase or andesine) porphyry compositionally equivalent to a granodiorite or quartz diorite; the term was introduced to distinguish a porphyry that contains plagioclase phenocrysts from one containing alkali feldspar phenocrysts.

Rhomb-porphyry—a porphyritic alkali syenite with phenocrysts of augite, occasional olivine, and anorthoclase or K-oligoclase (with rhombohedral cross-sections) in a groundmass composed mainly of alkali feldspars.

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Cross-references: *Granites and mineralization*; *Hypabyssal rocks*; *Lamprophyre*; *Nucleation of minerals in magma*; *Phenocryst*; *Textures of igneous rocks*.

PREFIXES, SUFFIXES

Prefixes

- | | |
|--------|---|
| Anchi | almost (anchimonomineralic). |
| Apo | implying derivation of one kind of rock from another, such as by metasomatic change without destruction of original texture, or by devitrification of a volcanic rock (apo-obsidian). |
| Blasto | used in the description of a metamorphic rock to indicate a relict texture (blastoporphyratic). |
| Eo | added to names of volcanic rocks to indicate alteration by devitrification or recrystallization (eorphylite). |
| Epi | signifying alteration (but not used consistently, e.g., epidiorite); also used in such terms as epibatholithic |

	(implying peripheral) and epizone (implying uppermost).
Hemi	signifies half (hemiclastite).
Holo	signifies whole (wholly) (holocrystalline).
Hyalo	signifying a glassy rock of corresponding chemical composition (hyalobasalt).
Iso	signifying equality (isotherm).
Kali	signifies an absence of plagioclase (or <5%) in an igneous rock (kalialaskite).
Leuco	indicates leucocratic character (leucogranite, leucosome).
Lith	stone or stone-like (lithic).
Melano-, mela-	indicates melanocratic character (melanosome, melasome).
Meta	indicating that the rock type has been metamorphosed (metasediment).
Micro	signifying a microscopic characteristic (e.g., graphic vs. micrographic texture) or a finer grain (e.g., granite vs. microgranite).
Ortho	indicates derivation from an igneous rock (orthogneiss, orthoamphibolite).
Palaeo-(paleo-)	indicates a pre-Tertiary age (or pre-Carboniferous in the case of some authors) and generally altered characteristics (cf. Palaeovolcanology).
Para	indicates derivation from a sedimentary rock (paragneiss, para-amphibolite).
Post	after (post-tectonic).
Pre	before (pre-tectonic).
Syn	together (syntectonic).
Vitro	indicates the presence of abundant glass (vitrobasalt).

Suffixes

Blastic	signifies a feature or texture formed entirely by metamorphism (porphyroblast, porphyroblastic).
Clasis (clastite)	signifies broken (holoclastis, holoclastite).
Lith	rock or rock-like (stromatolith, xenolith).

PREHNITE-PUMPELLYITE FACIES

The *prehnite-pumpellyite facies* is a very low-grade mineral facies intermediate between the zeolite and higher-grade facies, and is characterized in sandstones and metavolcanic rocks of appropriate composition by the presence of prehnite and/or pumpellyite in the absence of zeolites, actinolite, lawsonite, or

jadeite. Plagioclase is usually albitized, but relict calcic plagioclase may be preserved along with augite, hornblende, biotite, and orthoclase. Prehnite and pumpellyite tend to be sparse in psammites, but are more plentiful in metabasites. Spongy crystals of prehnite enclosing quartz are interpreted as resulting from the breakdown of laumontite or heulandite. Zeolites are entirely lacking except perhaps in later cross-cutting veins and joints.

Field occurrence

Thickened prisms of sedimentary and volcanic rocks accreted in present or former continental collision zones such as the circum-Pacific belt, the European Alps, W Appalachians, and the Caledonian belt in the British Isles (e.g., Bevins and Rowbotham, 1983), are particularly prone to this type of alteration. The facies is well represented in structurally simple burial metamorphic sequences and also in Precambrian greenstone belts in which the newly-formed assemblages are often particularly well developed in flow tops and scoriaceous lavas. Characteristic mineral assemblages have been reported less commonly from the present-day oceanic crust in which they have evidently been produced by active hydrothermal alteration involving modified sea water convecting through volcanic rocks under a steep thermal gradient. The upper portions of ophiolites commonly show similar alteration. Whether this results from sea-floor metamorphism, or is an overprint of later burial or Alpine-type metamorphism may be debatable. Evarts and Schiffman (1983) have argued that the metamorphic phenomena in the Del Puerto ophiolite of California are of sea-floor origin. The prehnite-pumpellyite facies tends to be poorly developed in near-surface on-land hydrothermal systems, and pumpellyite is commonly absent in such situations.

Prehnite-pumpellyite facies metamorphism of arenites and igneous rocks is usually not accompanied by penetrative deformation, although brittle-fracture phenomena with veining by quartz, prehnite, calcite, and other minerals, including axinite, are commonly conspicuous. Associated argillites may show weak to moderately-developed cleavage. Mineralogical reconstitution in argillites and limestones is less obvious than in other lithologies, although smectites have largely disappeared. Impure limestones may contain prehnite and pumpellyite.

Definition

Coombs (1960) proposed the recognition of a *prehnite-pumpellyite metagraywacke facies* to

include those mineral assemblages produced under physical conditions in which the following coexist: quartz–albite–prehnite–chlorite and quartz–albite–pumpellyite–chlorite, without zeolites, jadeite, or lawsonite. He provisionally suggested two zones or subfacies, one at lower grade in which prehnite is stable but the pair pumpellyite–actinolite is not, and one of higher grade in which pumpellyite–actinolite is stable, stilpnomelane becomes widespread and prehnite disappears. This usage, without the subfacies, was followed by Turner (1968). Hashimoto (1966) proposed that those rocks in which the mineral pair pumpellyite–actinolite is stable in coexistence with albite and chlorite be separated as the pumpellyite–actinolite schist facies, and that the prehnite–pumpellyite metagraywacke facies be correspondingly restricted. The name “prehnite–pumpellyite facies” in the restricted sense of Hashimoto has subsequently been used by many writers, including Turner (1981), and is adopted here.

Mineral assemblages

Typical assemblages include (Fig. 1):

1. quartz–albite–pumpellyite–prehnite–chlorite–sphene (titanite) (e.g., in many graywackes);
2. quartz–albite–epidote–prehnite–pumpellyite–chlorite;
3. albite–epidote–pumpellyite–chlorite–hematite–sphene (titanite) $\pm$ quartz (e.g., in some metabasites); and
4. quartz–albite–prehnite–andradite–grossular–chlorite.

Other possible phases include phengitic mica, K-feldspar, calcite, dolomite, celadonite, and axinite. It is to be emphasized that the occurrence of prehnite and/or pumpellyite is not in itself diagnostic of the prehnite–pumpellyite facies. Under conditions of very

low μ_{CO_2} and relatively high $a_{\text{Ca}^{2+}}/a_{\text{H}^+}$ in the coexisting fluid phase, prehnite can form at very low temperature, well within the fields of heulandite, laumontite, or wairakite, all diagnostic of the zeolite facies. Pumpellyite also overlaps these minerals.

Relation to contiguous facies

The transition from zeolite to prehnite–pumpellyite facies, as in southern New Zealand and the Kii Peninsula, Japan, is marked by the disappearance of zeolites, notably laumontite. In low-pressure facies series, as in the Tanzawa Mountains, Japan, the sequence with increasing grade of metamorphism is zeolite facies (often with wairakite), prehnite–pumpellyite facies, greenschist facies (Seki et al., 1969). In intermediate-pressure facies series, a zone of pumpellyite–actinolite assemblages intervenes between the prehnite–pumpellyite facies and the greenschist facies, as has been described on the flanks of a typical regional metamorphic terrane in southern New Zealand by Bishop (1972). In high-pressure facies series the prehnite–pumpellyite facies may pass into the lawsonite–albite–chlorite facies as occurs elsewhere in southern New Zealand (see **Zeolite facies**). The facies corresponds approximately to the anchimetamorphic zone of Alpine petrologists as defined by illite crystallinity (e.g., Kisch, 1983).

Physical conditions of metamorphism

Seki and Liou (1981) indicate temperatures for the prehnite–pumpellyite facies of about 210°C to 310–390°C at pressures of up to about 4 kb and at very low X_{CO_2} . They point out that the key minerals can be synthesized over a wider range of temperature and pressure than this. Liou et al. (1987) place the lower-pressure, higher-temperature parts of the field in the recently recognized *prehnite–actinolite facies*.

DOUGLAS S. COOMBS

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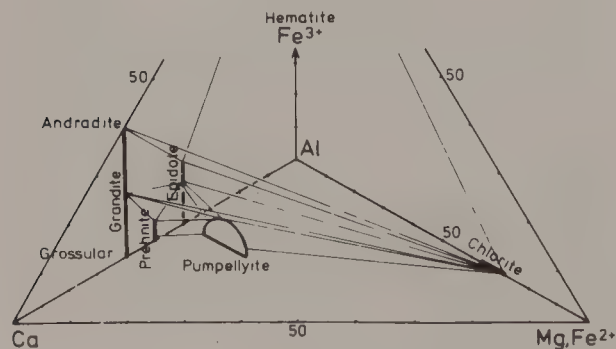


FIGURE 1. Some possible assemblages of the prehnite–pumpellyite facies in the system Fe_2O_3 – Al_2O_3 – CaO – $(\text{Mg}, \text{Fe}^{2+})\text{O}$ in the presence of excess SiO_2 and H_2O . (From Coombs et al., 1977.)

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Cross-references: *Burial metamorphism; Greenschist facies; Metamorphic facies; Ocean-floor metamorphism; Pumpellyite–actinolite schist facies; Regional metamorphism; Zeolite facies.*

PTYGMATIC VEIN

The term “ptygmatic” (from the Greek *πτυγμα* which means anything that is folded) was introduced by Sederholm (1907) for disharmonically folded pegmatitic material in migmatites in southern Finland. The most obvious characteristic of a ptygmatic vein (ptygma of some authors) is its tortuous shape (Fig. 1), which commonly exhibits an apparent shortening of up to sixfold. Axial planes of individual fold forms generally lack parallelism while the quartzofeldspathic layer that is usually <5 mm thick shows no apparent thickening in fold hinges or thinning in limbs. Wavelengths of fold forms of individual ptygmatic veins range from fairly constant to highly variable, with some having both highly tortuous and nearly planar portions. Although some adjacent veins may be nearly parallel, typically they are not. Suggested modes of origin include (1) spreading and/or laminar flow, (2) replacement–expansion flowage, (3) metasomatism, (4) filling of tortuous fissures, (5) aberrant injection, (6) tectonic deformation, (7) syntectonic injection,

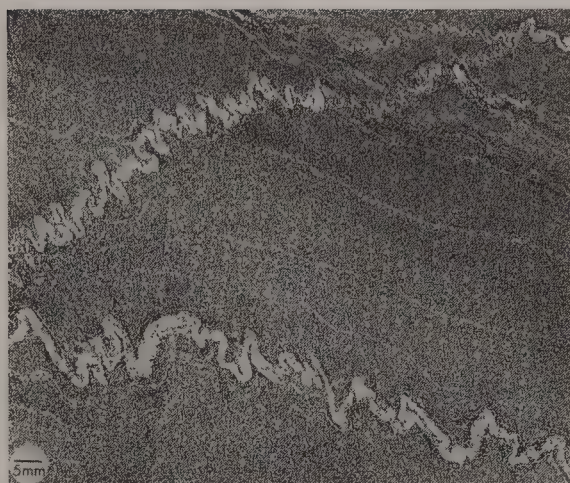


FIGURE 1. Ptygmatic veins composed almost entirely of quartz and feldspar in a biotite–quartz–feldspar host; injection zone of Alexandria batholith, New York State, U.S.A.

(8) volume reduction of host rock by squeezing, (9) plastic deformation, (10) magmatic flowage, (11) development within a passive host, (12) metamorphism of convoluted sedimentary layers, and (13) tectonic deformation followed by partial anatexis, localized magmatic flowage, and subsequent consolidation under static conditions (Bartholomew and Dietrich, 1973). Ptygmatic structures have been produced experimentally by the contraction of a competent band contained in a more ductile material (cf., Fig. 2) and Ramsay (1967) both relates their development to a particular state of finite strain during an irrotational progressive deformation and points out that their form resembles the curve known as an *elastica*.

Ptygmatic veins are common in many gneissic and migmatitic terranes (e.g., Read, 1931; Mehnert, 1968; Snowden, 1984). Their characteristic shapes permit distinction from other folds that deform migmatitic quartzofeldspathic material but that have consistently parallel axial

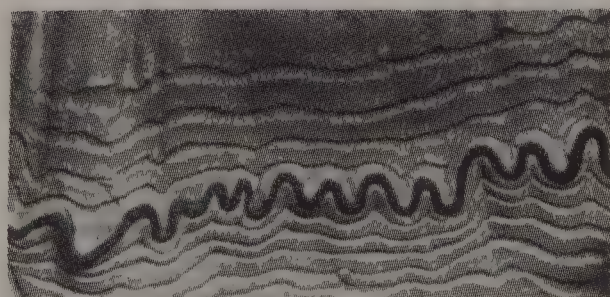


FIGURE 2. Ptygmatic structure produced experimentally by the contraction of a competent band contained in a more ductile material. (From Ramsay, 1967.)

planes and result from slip movement along foliation planes (Hopgood et al., 1976).

R. V. DIETRICH

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Cross-references: *Gneiss*; *Injection complex*; *Migmatite*; *Migmatite—origin of neosome*; *Migmatite—structural relationships*.

PUMPELLYITE-ACTINOLITE SCHIST FACIES

The pumpellyite-actinolite or *pumpellyite-actinolite schist facies* is a very low-grade mineral facies intermediate between the prehnite-pumpellyite and greenschist or glaucophane schist (blueschist) facies. It was first defined by Hashimoto (1966), although the existence of such a facies was suggested by De Roever (1947) on the basis of experience in Sulawesi, Indonesia. C. O. Hutton had earlier described typical assemblages from southern New Zealand. Rocks of this facies occur over quite large areas in the low-grade parts of many metamorphic belts characterized by intermediate to moderately high-pressure facies series such as in New Zealand, Japan, and the Alps (e.g., Coombs et al., 1976). The diagnostic mineral assemblage is here accepted as pumpellyite-actinolite-chlorite, without glaucophanic amphibole. A typical mineral assemblage in metagraywackes and some metabasites is quartz-albite-pumpellyite-actinolite-chlorite-phengitic muscovite-sphene(titanite) $\pm$ epidote $\pm$ stilpnomelane. Grandite garnet is occasionally found coexisting with chlorite under conditions of very low μ_{CO_2} . In the case of pelitic schists, it

may not be easy to distinguish pumpellyite-actinolite schist facies assemblages from those of the greenschist facies.

Bishop (1972) has mapped pumpellyite-actinolite schist facies rocks, 9 km thick, intervening between the prehnite-pumpellyite and greenschist facies in the Dansey Pass area of New Zealand. Rocks of the pumpellyite-actinolite facies in such terranes are typically schistose, but Bishop shows that "isotects" defining progressive changes of fabric from massive, nonfoliated graywacke, through schistose graywacke, to schist, are not strictly parallel to isograds separating the mineral facies. The prehnite-out isograd was used to define the beginning of the pumpellyite-actinolite facies, but this excludes some rocks containing the critical assemblage defined above.

Where the facies is developed in rather lower-pressure facies series with correspondingly higher temperature gradients, as in the burial metamorphic sequence of the Proterozoic Hamersley Basin of W Australia (Smith et al., 1982), prehnite is an important phase in rocks of appropriate composition. In such suites, in contrast to those of higher-pressure facies series, it persists to higher grades than pumpellyite, and the *prehnite-actinolite facies* intervenes between zeolite or prehnite-pumpellyite facies and greenschist facies (Liou et al., 1985).

Nakajima et al. (1977) show that the following reaction may divide the facies into lower- and higher-grade parts: ferrian pumpellyite + hematite + quartz = chlorite + epidote + actinolite + water. At the lower grades, pumpellyite, chlorite, and actinolite can thus coexist with hematite for appropriate bulk compositions, and at higher grades, with epidote. Nakajima et al. have modeled changing compositions of the epidote with rising temperature. The range of bulk compositions in which pumpellyite can occur is progressively restricted, and it is finally eliminated.

The reaction pumpellyite + chlorite + quartz = epidote + actinolite + H_2O at $P_{\text{H}_2\text{O}} = 7$ kb is $370 \pm 20^\circ\text{C}$ (Nitsch, in Bishop 1972) and the breakdown of pure Mg-Al pumpellyite to clinozoisite, garnet, and chlorite is located close to 380°C at the same pressure by Schiffman and Liou (1980), with steep positive slope. These data constrain upper temperature limits for the facies. Pumpellyite-actinolite facies rocks have been recognized up-grade of the anchizone-epizone boundary, as defined by illite crystallinity in the Swiss Alps. From consideration of fluid inclusion data as well as coal rank reflectivities, Frey et al. (1980) consider that minimum temperature and pressure for this boundary are not less than 270°C and 1.2 kb,

respectively. Temperatures and pressures for typical representatives of the facies may thus be in the general region of about 300–350°C at pressures of several kilobars. (For a schematic representation of the position of the pumpellyite–actinolite facies in the petrogenetic grid for low-grade mineral facies, see **Zeolite facies**.)

DOUGLAS S. COOMBS

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- Cross-references: *Burial metamorphism*; *Glaucophane schist facies*; *Greenschist facies*; *Metamorphic facies*; *Metamorphic facies series*; *Prehnite–pumpellyite facies*; *Schist*; *Zeolite facies*.

PYROCLASTIC ERUPTIONS

Explosive or pyroclastic volcanic eruptions differ from lava eruptions in that the lava is broken into solid fragments or liquid clots, called volcanic ejecta, by the escape of gases contained in the lava. As used here the term

pyroclastic eruption excludes phreatic eruptions, which result when hot lava comes into contact with water.

Pyroclastic eruptions can be divided into normal explosive eruptions and abnormal explosive eruptions, on the basis of the energy released and the volume of material erupted per unit time. Thus, an abnormal eruption releases some 10^{26} joules of thermal energy in the space of a few days; a normal eruption releasing this same amount of energy would have to continue for several years. Eruption clouds (suspensions of fragmental material in rising gases) produced during normal eruptions usually do not reach heights of much more than 1,500 m, whereas the eruption cloud of an abnormal eruption may reach 15,000 m or more. Normal eruptions are commonly associated with lava eruptions, whereas abnormal eruptions are not, although they may be followed by extrusive activity.

Normal explosive eruptions will be discussed in order of increasing viscosity of the lava involved. The more viscous the lava, the more intermittent is the escape of gases from it, because a greater gas pressure must be built up for the gas to escape from a viscous lava than from a fluid one. This difference in viscosity may in turn result from differences in the composition, temperature, or crystallinity of the lava: high SiO_2 content, low temperature, and high crystal content all increase the viscosity of a lava. As a result, the type of explosive activity commonly changes during the course of a given eruption.

If the lava is very fluid, the contained gases can escape rather easily and continuously. During a *Hawaiian-type* pyroclastic eruption they may throw fountains of molten basaltic lava to heights of a few hundred meters or more for hours on end. The ejecta formed during such eruptions depend on the gas pressure involved. Low pressures produce low fountains that eject rather large plastic clots that tend to weld together on impact to form dribble and spatter cones. Higher fountains produce much finer ejecta, and the droplets usually solidify before striking the ground; their ejecta are of two general types: vesicular lapilli (scoria) and ash, and threads of glass (Pele's hair). The ash and lapilli pile up around the vent to form an ash or scoria cone, while the Pele's hair is usually carried considerable distances downwind.

With increasing viscosity, gas release becomes spasmodic rather than continuous. The lava may still be ejected as plastic clots but these are rigid enough when they strike the ground to retain the shape given them during their flight through the air. Such activity is of the *strombolian type*. The pyriform bombs produced result from the

break-up of ribbons of ejected lava that rarely survive intact. In addition, finer ejecta of lapilli and ash size are formed in large quantities, though these are less vesicular than the analogous ejecta of hawaiian eruptions. These ejecta for the most part pile up around the vent to form an ash or cinder cone.

With still higher viscosity eruptions of the *vulcanian type* occur. This activity also is spasmodic, but the viscosity of the lava here is so great that the material is ejected as fragments rigid enough to resist modification in shape by their flight through the air or by their impact. The larger ejecta are mainly blocks rather than bombs. Many have a well-developed bread-crust surface, with a network of fissures up to 1 cm deep that develop by vesiculation and expansion of the interior after a rigid crust has formed. Most ejecta, however, are slightly vesicular lapilli and very abundant ash, and these eruptions are characterized by dark convoluted eruption clouds. The lapilli and coarse ash fall near the vent to form a cinder cone, while the finer ash may be carried some distance downwind.

Eruptions of the *peléan type* are probably not explosive or pyroclastic in the usual sense. They are characterized by two or three phenomena that actually appear to be side-effects of the development of an intrusion dome in the summit crater or on the upper flanks of a stratovolcano. In order of decreasing frequency of occurrence, these include the following features.

1. *Agglomerate flows* are suspensions of incandescent solid fragments in hot expanding gases that result from the falling away of portions of the flank of a growing dome; under the influence of gravity, these rush down the slopes of the cone, reaching speeds of >100 km/hour; they are one version of a more general phenomenon called a *nuée ardente* (examples—Mt. Pelée, Martinique, in 1902 and 1929; Merapi, Java, last in 1942–1943; and Santiaguito, Guatemala, last in 1973).

2. *Directed blasts* are produced by the sudden release of pressure on the magma in the conduit and reservoir of the volcano following a rather long period (100–250 years?) of quiescence (examples—Mt. Pelée (1902); Lassen Peak, California, U.S.A. (1915); Mt. Lamington, Papua–New Guinea (1951); and Mt. St. Helens, Washington State, U.S.A. (1980)).

3. *Theater collapse* is caused by the collapse of a considerable portion of the dome, resulting in the development on the dome of a cirque-like collapse theater and in front of it of a hummocky avalanche deposit; no well-documented examples have occurred in historic

times—probably the best-preserved prehistoric example is that on Chaos Crags in N California, U.S.A.

The abnormal explosive eruptions are much less common than the normal ones. As a result, their classification and characterization is imperfect, but large amounts of thermal energy are released and large amounts of ejecta erupted over short time spans.

Plinian- and *katmaian-type* eruptions appear to represent the two end-members of one type of abnormal explosive activity. Such activity is preceded by a long period of quiescence (usually on the order of 500 years), during which the magma in the conduit or upper lava reservoir becomes gas-saturated and a separate gas phase may be produced. Release of pressure by failure of the conduit plug or of the cone itself results in the rapid and thorough-going vesiculation of this lava, and in its ejection a mass of still vesiculating pumice lumps and glass shards. With increase in the diameter of the orifice, this ejection changes from jet-like (*plinian*), the eruption cloud climbing rapidly to 15,000–30,000 m, to the simple “boiling over” of the volcano (*katmaian*), to form pyroclastic flows of hot fragments in expanding gases that pour down the sides of the cone. Intermediate situations are common in which the finest ejecta are carried to great heights at the same time that the heavier fragments are sweeping down the side of the cone.

Deposits of *plinian* origin have an elongate distribution pattern (Fig. 1) that is controlled by the wind direction in the upper atmosphere. The heavier and larger particles fall out first and accumulate near the vent to form a thick chaotic accumulation of crystals and pumice lumps; away from the vent the deposit thins and becomes a finer-grained, better-sorted, and a finely layered accumulation of glass shards and dust (examples—Vesuvius (79 A.D.), Tambora (1815), Krakatau (1883) and Agung (1963), Indonesia, Arenal, Costa Rica (1968) and El Chicon, Mexico (1982)).

Deposits of *katmaian* origin are stream-like in plan view, their distribution being tightly controlled by the stream valleys on the sides of the cone. Throughout their length the deposits are an unsorted mixture of pumice lumps, glass shards, and crystal fragments. The escape of entrapped gases may go on for years, and the action of these rising vapors may consolidate the deposit by crystallization of and between the glass shards. A “pure” *katmaian* eruption apparently occurred on Cotopaxi Volcano, Ecuador, in 1877, but most other historic *katmaian* eruptions also involved *plinian*

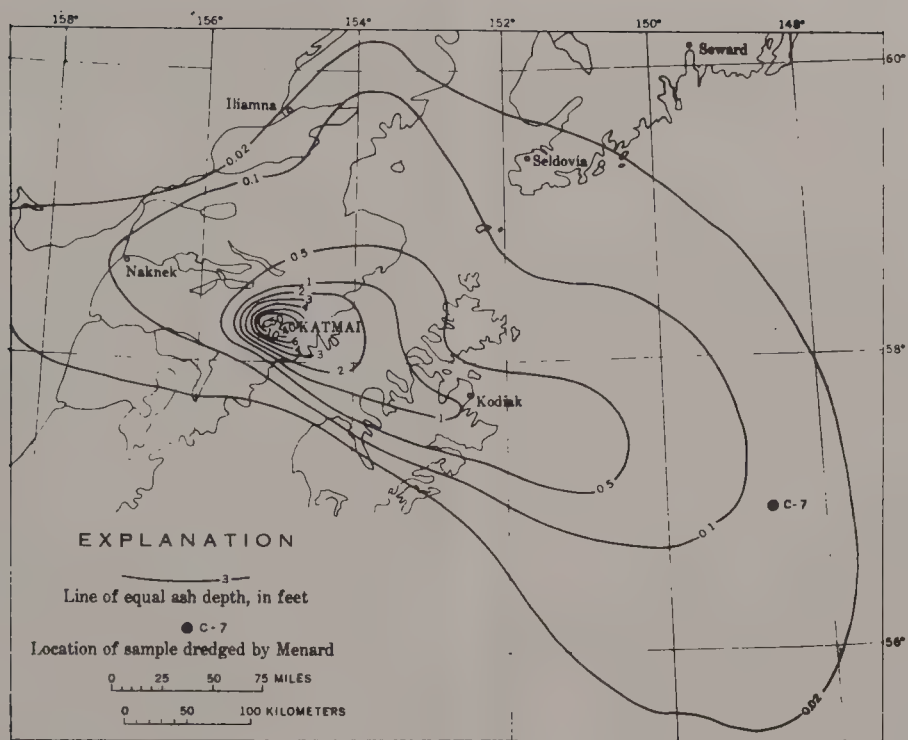


FIGURE 1. Map of Mount Katmai and the Valley of Ten Thousand Smokes, Alaska, U.S.A., showing thickness of ash deposit from the eruption of June 1912.

activity (Katmai, Alaska (1911), Bezmyannyi (1955), Mayon, Philippines (1968)).

Eruptions of another abnormal type, here called the *rotoruan type*, have never been described by eye-witnesses, so that their nature must be reconstructed from their products. They are characterized by the massive eruption from one or more fissures of a suspension of gas-charged pumice, glass shards, and crystal fragments in hot expanding gases. The internal viscosity of the suspension must be very low, for its deposits are spread as a more or less uniform sheet over areas that may measure thousands of kilometers. Over its entire extent the sheet shows little variation in composition, grain size, sorting, or stratification; the last two features are invariably absent. Often the temperature of the fragments when deposited was high enough to allow them to weld together, at least in the lower parts of the deposit. *Ignimbrites*, the products of such eruptions, now cover large areas in the western U.S.A., on the North Island of New Zealand, on Sumatra, in Armenia and in E Siberia, U.S.S.R. The removal of such large amounts of material results in large-scale collapse of portions of the upper crust, forming volcano-tectonic depressions.

CHARLES P. THORNTON

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Cross-references: *Calderas and volcano-tectonic depressions*; *Endogenous dome*; *Explosive volcanic eruptions—classification*; *Nuée ardente*; *Pyroclastic flows*; *Tephra*; *Volcanoes and volcanism*.

* Gives extensive reference list.

PYROCLASTIC FLOWS

The term *pyroclastic flow* is generally applied to the units of volcanic rock that have resulted from *nuées ardentes*. It could be argued that it should be enlarged to include any pyroclastic deposit that has been moved by a flow mechanism, for example, the base-surge deposits associated with phreatomagmatic basic volcanicity (Waters and Fisher, 1971). However, the term is still used in its original restricted sense. The term *ash-flow tuff* is used for tuffs deposited from a mobile gaseous cloud or ash flow. The rock is a consolidated but not necessarily welded deposit. Some authors use the term as synonymous with *ignimbrite*, a term that originally implied dense welding but which now is not used in such a restricted way. Other

authors use the term ash-flow tuff as a type of ignimbrite. These rocks are most frequently of rhyolitic, dacitic or andesitic composition, although small flows of basic material have been observed.

The magmatic material in a pyroclastic flow normally cools into a volcanic glass. If a considerable thickness of material accumulates quickly enough (either from a single extrusion or by a rapid succession of flows), this glassy material will flatten in the middle part of the accumulation, to give welded tuff between the upper and lower layers of more rapidly cooled, unwelded material. In the absence of such welded portions the pyroclastic flow origin may not be immediately apparent. It would, however, be suggested by the absence of bedding in a glass-rich pyroclastic rock. Proof of the pyroclastic flow origin of such rocks might only be established by tracing them laterally into welded tuffs.

Welding generally results from the pressure of overlying material pushing the glassy constituents of the rock together while they are still hot and plastic. A rapid succession of pyroclastic flows would allow the accumulation of later upper flows before the earlier underlying flows had cooled. Hence in welded pyroclastic flows, cooling units are of greater importance in determining the characters of the rocks than are the individual flows: indeed the joins of flows are frequently very difficult, if not impossible, to trace through densely welded tuffs. They may be marked by a thin parting of ash-fall tuffs or reworked (bedded) parts of the upper surface.

Zoning of cooling units

Cooling units may be several hundreds of meters in thickness, vertically zoned, and have partially welded tuffs between the nonwelded and densely welded rocks (Fig. 1). The welding together of the numerous fragments of volcanic glass is accompanied by flattening with the boundary between the nonwelded and partially welded portions of devitrified tuffs being shown

by the first sign of such flattening. Collapsed pumice generally shows progressive elimination of pore spaces (vesicles), which may form up to 75% volume in uncollapsed rock. The development of a completely nonporous rock may, however, be prevented by the presence of lithophysae, which can occur in any part of a cooling unit.

The degree of collapse in any part of a pyroclastic flow depends on (1) plasticity, which is closely related to temperature, and (2) the pressure exerted by overlying material (lithostatic load). Hence the position within a cooling unit of the zone of maximum distortion may vary in relation to the thermal history of the unit. If the whole unit accumulated very quickly by a massive single extrusion of material, or by numerous flows of similar temperatures in very rapid succession, the initial temperature distribution would be fairly uniform, lithostatic load would be the dominant factor and maximum flattening would occur near the base of the unit. If, however, the building up of the lower portion of the unit were more gradual, some cooling would have occurred before the maximum lithostatic load was applied: higher temperature plus smaller load at a higher level in the unit would then result in greater compaction than in the part where the temperature was lower and the load higher. Some cooling units show a more complex pattern of zoning, possibly due to periodicity of eruption, unequal area of distribution of individual flows or marked differences of the temperature of flows. A simple cooling unit may grade laterally into a compound cooling unit and eventually into two separate units.

The flattened lumps of pumice within a welded tuff are often best shown on partially weathered surfaces, especially in a dense vitrophyre, where in the fresh rock there is no color contrast between the fragments of pumice and the glassy matrix. Here partial weathering lightens the matrix before the pumice fragments are affected. In the compaction of less densely welded material, however, the darkening of pumice fragments precedes the darkening of the matrix of glass shards. Hence in some rocks a similar color contrast may be visible on fresh surfaces. Such a rock containing dark lenses of glass in a lighter matrix is known as *piperno* (originally a local Italian name applied to a group of eutaxitic trachytic tuffs) and the glass lumps are known as *fiamme*.

The flattening of the pumice fragments tends to accentuate initial irregularities to give frayed ends in sections normal to the flattening. The overall shape of such collapsed fragments is normally disk-like, but they may occasionally be tongue-shaped. This is thought to be due to flow

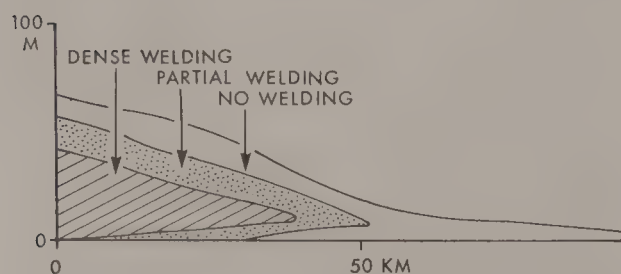


FIGURE 1. Schematic representation of the typical zonal arrangement of a thick cooling unit of pyroclastic flows; vertical scale much exaggerated.

movement in the compacted or compacting mass, the most probable cause being differential compaction over buried topography. Another possibility is fault movement in the area of the flow. The flattening of a matrix of glass shards to give a eutaxitic texture may be partly visible in the hand-specimen but is best seen by microscopic examination (Ross and Smith, 1961, figs. 20–36 and 38–41).

Although flattening of pumice to give *fiamme* is usually associated with thick cooling units and large lithostatic load, this has been seen in thin pyroclastic flows: Gibson and Tazieff (1967) recorded a very thin (0.3–20 m) pyroclastic flow in which very hot, plastic lumps of pumice appear to have flattened immediately even in the thinnest part of the sheet and to have vesiculated to give spherical bubbles after flattening. This flow became welded throughout.

Crystallization of the originally glassy constituents of pyroclastic flows can be classed as primary if it occurred soon after the emplacement of the flow, but secondary if it is of considerably later origin; possible patterns of primary crystallization have been described by Smith (1960). Devitrification of glassy constituents into spherulitic or axiolitic growths of cristobalite and feldspar may be confined to a narrow interior zone of a cooling unit. In thicker units it is likely to have affected all but the lower selvages of unwelded and poorly welded tuffs and the lowermost part of the densely welded material. A vapor-phase crystallization may affect the upper poorly-welded or nonwelded portions of a unit during or after devitrification; this leads to a coarser-grained aggregate of crystals growing in pore-spaces. A granophyric type of crystallization (micrographic intergrowth of quartz and feldspar) is sometimes encountered in the central parts of very thick cooling units. Secondary devitrification is revealed by an absence of any relationship to the vertical zoning of the flow.

While axiolitic crystallization may retain the original eutaxitic textures of a welded tuff, the other types of crystallization are more likely to obliterate and obscure the original characters of the rocks involved. Photomicrographs of devitrification textures are given by Ross and Smith (1961).

Columnar jointing is common in the welded portions of pyroclastic flows but generally absent in the unwelded or poorly welded portions. Fenner (1948) also described columnar jointing in very porous pyroclastic flow rocks, which had become indurated by crystallization; those are referred to as *sillars*. It has been suggested that this term should be used for any nonwelded rock of pyroclastic flow origin, but it is probably better reserved for indurated and

largely crystalline rocks. From microscopic examination Smith (1960) also recognized incipient welding in these rocks, but no large degree of flattening or welding can be allowed in this term.

Origin

The origin of thick cooling units of pyroclastic flows presents some difficulties, since *nuées ardentes* of historic times have generally produced either thin accumulations which show no sign of welding (Mt. Pelée), or moderately thick flows which are only slightly welded, like the Novarupta (Katmai) 1912 flow averaging rather less than 30 m thick. In December 1951 Hibokhibok in the Philippines filled the lower parts of a valley with a pyroclastic flow more than 120 m thick, but erosion has not yet progressed sufficiently to reveal the degree of flattening in these rocks. Moreover the volume of rock ejected by this explosion from the base of a volcanic dome was only a fraction of that involved in some prehistoric cooling units. Overflowing *nuées ardentes* from elongate fissures or ring fractures is the most commonly postulated origin for such voluminous cooling units.

There is frequently an association of mudflow deposits (*lahars*) with pyroclastic flows. These are normally confined to the valleys which descend from the distal parts of the flows (Francis, 1976; Decker and Decker, 1981).

A normal magmatic flow origin has been advocated for some pyroclastic flows: such an origin is associated with the term “*tufflava*”. In recent years it has been extensively advocated only in the U.S.S.R., but the 1961 Moscow symposium (English version—Cook, 1966) reveals considerable disagreement amongst Soviet geologists on the origin, relative importance, or even the existence of *tufflavas*.

Studies of the deposits of recent volcanic eruptions, such as that of Tambora, Indonesia in 1815 (Self et al., 1984), have revealed much information about the nature and origin of pyroclastic flows. Much has also been learnt from the products of Cenozoic subaerial volcanic activity in the western United States of America, such as those of the 600 km³ (dense-rock equivalent) Bishop Tuff, California (Hildreth, 1979). This consists of a basal plinian air-fall tuff and a group of overlying ash-flow tuff sheets from which Anderson et al. (1989) have found evidence of pre-eruption concentrations of volatiles in separate parts of the parent magma body.

Pyroclastic flows were at one time regarded as being indicative of terrestrial conditions. However submarine unwelded Ordovician tuffs have

been recorded in North Wales, U.K. (Howells et al., 1973). These are rich in shards and pumice and grade laterally into welded tuffs that are also considered to have been formed in a submarine environment and to have moved across the sea floor as a hot, gas-suspended flow that was largely insulated from the surrounding water by a carapace of quenched material.

W. E. TREMLETT

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Cross-references: *Acid(ic) igneous rocks*; *Columnar jointing*; *Explosive volcanic eruptions—classification*; *Nuée ardente*; *Pyroclastic eruptions*; *Tephra*; *Textures of igneous rocks*; *Volatiles*; *Volcanic glass*; *Volcanoes and volcanism*.

PYROLITE

The term *pyrolite* has received wide acceptance as a useful term for a model composition for Earth's upper mantle. In its simplest form

the term embodies the hypothesis that the upper mantle is mineralogically dominated by pyroxenes and olivine (hence *pyr-ol-ite*) but is capable of yielding basaltic magmas by partial melting leaving residual more-refractory peridotite (olivine + pyroxene(s)) (Ringwood, 1962). The term is a general one and was coined deliberately to avoid use of established names for natural rocks such as dunite, peridotite, harzburgite, lherzolite, saxonite, websterite, or wehrlite. Some natural lherzolites match the model pyrolite in mineralogy and chemical composition and are derived from the upper mantle by tectonic or magmatic transport.

Following improved understanding of the petrogenesis of basalts, in which it was shown that parental Hawaiian olivine tholeiite magma could be in equilibrium with harzburgite residue (olivine + orthopyroxene) at depths of 60 km, Ringwood (1966) calculated a revised "Hawaiian pyrolite" composition by combining Hawaiian olivine tholeiite and harzburgite in the proportions 1:3. The choice of the 1:3 proportions was guided by the similarity in Mg, Fe, Si, Ca, and Al of the resulting composition to samples of natural lherzolites occurring in high-temperature peridotite intrusions, in inclusions in basalts, and garnet lherzolite inclusions in kimberlite pipes. With compelling evidence that the trace-element characteristics of mid-oceanic ridge basalts (depleted in large ion lithophile (LIL) elements) and "Hawaiian" or "within-plate" and rift valley basalts (enriched in LIL elements) required different source pyrolite compositions, Green et al. (1979) calculated a model pyrolite for the source regions of mid-ocean ridge basalts (Table 1).

The "Hawaiian" pyrolite composition of Table 1 has been extensively investigated for its melting relations, mineralogy as a function of pressure, temperature, and water content, and suitability in terms of physical properties and chemical composition as the mean composition of the upper mantle in those regions acting as sources for basaltic magma of Hawaiian type (tholeiite to olivine melilitite) (Green, 1973). While recognizing that it is arbitrary to calculate specific pyrolite compositions to presume that the upper mantle is not compositionally homogeneous, particularly in trace elements, even within or beneath the low-velocity zone, the model compositions of Table 1 obey both the geophysical and geochemical constraints for the upper mantle (Green and Liebermann, 1976). The melting behavior of the refractory Tinaquillo lherzolite composition has also been studied to high pressures, demonstrating a close similarity to the "Hawaiian" pyrolite composition.

D. H. GREEN

TABLE 1. Compositions of model pyrolite and similar natural rocks

	1	2	3	4
SiO ₂	44.7	45.0	45.2	45.0
TiO ₂	0.08	0.17	0.7	0.17
Al ₂ O ₃	3.2	3.9	3.5	4.4
Cr ₂ O ₃	0.45	0.41	0.4	0.45
Fe ₂ O ₃	0.1	—	0.5	—
FeO	7.5	8.3	8.0	7.6
MnO	0.14	0.14	0.1	0.11
NiO	0.26	0.24	0.2	0.26
MgO	39.8	38.0	37.5	38.8
CaO	3.0	3.5	3.1	3.4
Na ₂ O	0.18	0.40	0.6	0.40
K ₂ O	0.02	—	0.13	0.003
P ₂ O ₅	0.04	—	0.06	0.03

1. Tinaquillo lherzolite, Venezuela, a high-temperature diapiric lherzolite intrusion (Green, 1963).

2. Least refractory spinel lherzolite inclusion in basanite, Lake Bullenmerri, Victoria, Australia (Nickel and Green, 1984).

3. "Hawaiian" pyrolite model composition (Ringwood, 1966).

4. "MORB" pyrolite model composition (Green et al., 1979).

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- Cross-references: *Basalt*; *Harzburgite*; *Igneous rocks—tectonic setting*; *Kimberlite*; *Lherzolite*; *Mafic and ultramafic inclusions in igneous rocks*; *Melilitite*; *Upper mantle—experimental petrology*.

PYROMETAMORPHISM

Pyrometamorphism refers to changes induced in rocks by extremely high temperatures at low pressures. It is typically an effect noted in contact metamorphic aureoles immediately

adjacent to very hot, usually basic, high-level intrusions, and in xenoliths. Partial (or rarely complete) melting may result, leading to the formation of glassy rocks termed *buchites*. A typical buchite consists of a glass whose composition is close to that of the low- P_{H_2O} eutectic in the system quartz–albite–K-feldspar (Butler, 1961; Wyllie, 1961), containing euhedral or subhedral grains of tridymite (high-temperature SiO₂ polymorph), cordierite, mullite (an aluminosilicate), corundum, K-feldspar, and spinel.

Mineral assemblages are consistent with those of the pyroxene hornfels and sanidinite facies (see **Facies of thermal metamorphism**), probably reflecting temperatures in the range 850–1100°C, with pressures < 2 kb. At higher pressures, the resultant rock resembles the products of metamorphism in the lower-pressure cordierite granulite facies (see **Granulite facies**). Relics of lower-temperature parageneses are commonly preserved, and both phase and textural disequilibrium features are common.

In pelites, dehydration (marked by mica breakdown) is followed progressively by sanidine replacing microcline or orthoclase, mullite replacing sillimanite, and cordierite breaking down to hypersthene and spinel. Progressive melting, in which water, Na, K, and Ca preferentially enter the liquid (represented by glass), leaves a refractory residue that is gradually less siliceous and enriched in Fe, Mg, Ti, and Al. Buchites consisting of corundum–spinel–glass, mullite–glass, and spinel–mullite–glass may represent an extreme case.

In calcareous rocks, temperature, silica

activity (a_{SiO_2}), and CO_2 fugacity control the occurrence of a bewilderingly large range of mixed silicate-carbonate minerals. Silica-deficient calc-silicates, like melilite group members rankinite, larnite, and merwinite, occur with partly carbonated species like spurrite and tilleyite; most appear unique to this paragenesis. Decarbonation and higher silica activity result in such minerals as monticellite, forsterite, wollastonite, and hypersthene forming, while decarbonation and very low silica activity may result in minerals like brucite, periclase, and spinel forming.

Dehydration, decarbonation, and melting may radically change the composition of a rock. Siliceous limestones appear to be particularly prone to this, producing skarns (tactite if Fe-rich). The process operative in these cases should be termed *pyrometasomatism*.

J. ROBERT BUTLER

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Cross-references: *Assimilation*; *Aureole*; *Buchite*; *Contact (thermal) metamorphism*; *Facies of thermal metamorphism*; *Granulite facies*; *Pyrometasomatism*; *Skarn*.

PYROMETASOMATISM

The term *pyrometasomatism* is used in two different ways. The first usage applies to chemical changes that occur during pyrometamorphism when volatiles are mobile. Commonly, there is also evidence for mobility of constituents as Si, K, and Na (Butler, 1961). The second usage is that of Lindgren (1933). His pyrometasomatic deposits are a class of mineral deposits formed at relatively high temperatures near the contact zones of intrusive rocks and result from the replacement of enclosing rock by emanations from the intrusion. In this classification, pyrometasomatic deposits include a wide range of sulfide,

magnetite, gold, cassiterite, scheelite, and graphite deposits, in many cases associated with skarns.

J. ROBERT BUTLER

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Cross-references: *Contact metasomatism*; *Metasomatism*; *Pyrometamorphism*.

PYROXENITE

A *pyroxenite* is an ultramafic plutonic rock consisting of >60% pyroxene (either clino- or orthopyroxene, or both). Other phases include olivine, hornblende, micas, garnet, feldspar, spinel, ilmenite, and sulfides. They have a gradational relationship with peridotites.

The root names reflect the proportions of ortho- and clinopyroxene, and olivine; *orthopyroxenite* has >90% orthopyroxene, *clinopyroxenite* has >90% clinopyroxene, and *websterite* is a 2-pyroxene rock with <10% olivine. When olivine makes up 10–40% of the rock, these three root names are qualified with the prefix “olivine”. Likewise, the presence of feldspar (calcic plagioclase—always <5%), hornblende, garnet, and spinel in greater than accessory proportions is similarly indicated by a mineral prefix, e.g., garnet websterite, spinel orthopyroxenite, etc.

Pyroxenites may also be named according to the particular pyroxene species present, e.g., *augite* or *augitite*, *bronzitite*, *diallagite*, *diopsidite*, *enstatitite* or *enstatolite*, *hypersthene* (originally defined as a synonym of norite), where such a phase constitutes >90% of the rocks.

The following are other varieties of pyroxenite.

Aegirinolith—an aegirine-augite-sphene-magnetite pyroxenite.

Anabohitsite—an olivine-bearing pyroxenite containing hornblende and hypersthene with a high proportion of magnetite and/or ilmenite (Anabohitsy, Malagasy).

Bahiaite—contains orthopyroxene, amphibole, olivine, and a small proportion of ceylonite (Bahia, Brazil).

Bebedourite—contains biotite and accessory perovskite, apatite, and titanomagnetite.

Boderite—a nosean-sanidine-biotite augite. (Eifel, W Germany).

Canaanite—composed mainly of large white crystals of diopside (Canaan, Connecticut, U.S.A.).

Cromaltite—acmite-augite is the predominant mineral, with melanite as a characteristic component (Cromalt Hills, NW Scotland).

Ehrwaldite—an augitite containing both orthopyroxene and clinopyroxene.

Marchite—contains enstatite and diopside.

Niklesite—contains diopside, enstatite and diallage.

Salitrite—an aegirine-augite pyroxenite rich in sphene (Salitre Mtns., Brazil).

Sphenitite—a nepheline-bearing sphene pyroxenite.

Vibetoite—a biotite-amphibole pyroxenite with some calcite.

The occurrence of pyroxenites is similar to that of peridotites. They are present in nodules in basalts, nephelinites, and kimberlites, in the ultramafic mantle portions of ophiolite complexes, and as cumulate layers in basic igneous intrusions.

ADRIAN F. PARK

Cross-references: *Alkaline rocks—unsaturated; Intrusion—layered; Kimberlite; Mafic and ultramafic inclusions in igneous rocks; Ophiolite; Peridotite; Plutonic rocks—classification and nomenclature; Ultramafic rocks.*

R

RAPAKIVI TEXTURE

The term “rapakivi” was mentioned as early as 1694 by Urban Hjärne (Eskola, 1930) to refer to the feldspathic rubble in the weathered zone above certain granites in S Finland (rapakivi: Finnish, crumbly stone). These granites are remarkable for a particular texture, and the terms “rapakivi” texture and “rapakivi” granite were established by Sederholm (1891). In the strict sense they are characterized by a distinctive texture with three principal characteristics (Figs. 1, 2): (1) ovoidal alkali feldspar phenocrysts >3 cm in diameter; (2) mantling of some of these phenocrysts by oligoclase-andesine shells, up to 1–3 mm thick; and (3) the ubiquitous occurrence of two generations of alkali feldspar and quartz with the older idiomorphic quartz having initially crystallized as the high polymorph. Where the mantled phenocrysts outnumber unmantled ones, the rock is termed *wiborgite*; where the converse is true the rock is termed *pyterlite*. In both types the phenocrysts contain so-called concave quartz inclusions (Fig. 2) that frequently define concentric rings. In *wiborgite* ovoids a repetitive plagioclase mantling is often seen, producing an orbicular texture; the full range of diversity of these textures is described by Vorma (1971). In the broader sense, rapakivi texture is used where K-feldspar phenocrysts are mantled by plagioclase.

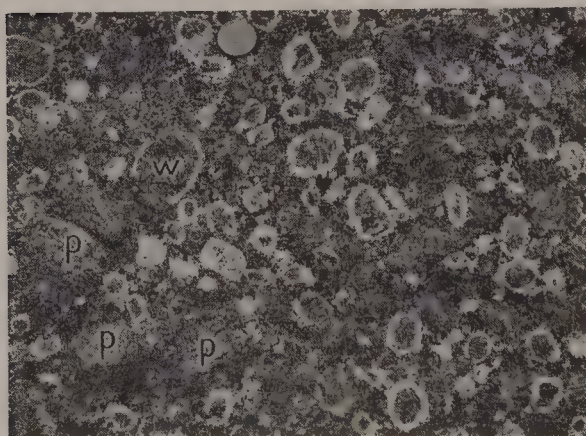


FIGURE 1. Rapakivi texture in wiborgite, Wiborg massif, SE Finland: w, wiborgitic ovoid; p, pyterlitic ovoid; $\times 0.1$.

The type rapakivi granites of S Finland are part of a discrete province that stretches from the Ukraine to S Norway; all the plutons are postorogenic with respect to the Svecokarelian orogeny (ca. 2.0–1.8 Ga) and were intruded during the period 1.7–1.5 Ga (Fig. 3). The rapakivi granites vary considerably in mineralogy and composition, although most fall in the fields of granite and alkali-feldspar granite in the QAP classification (Streckeisen, 1976), but some fall in the fields of syenite, quartz syenite, monzonite, and granodiorite. The largest masses (e.g., Wiborg massif, SE Finland and the Korosten complex, Ukraine) form parts of an igneous suite including other alkali-feldspar granites, granites, gabbros, and syenites with minor anorthosites and charnockite suite representatives, principally mangerite (Savolahti, 1956; Vorma, 1976). This association with charnockite suite rocks and anorthosites in the age range 1.8–1.5 Ga is reported from most of Precambrian shield areas, and is related to postorogenic activity (e.g., Scandinavia and Greenland) or intracratonic rifting (e.g., N America)—see review and references in Wind-

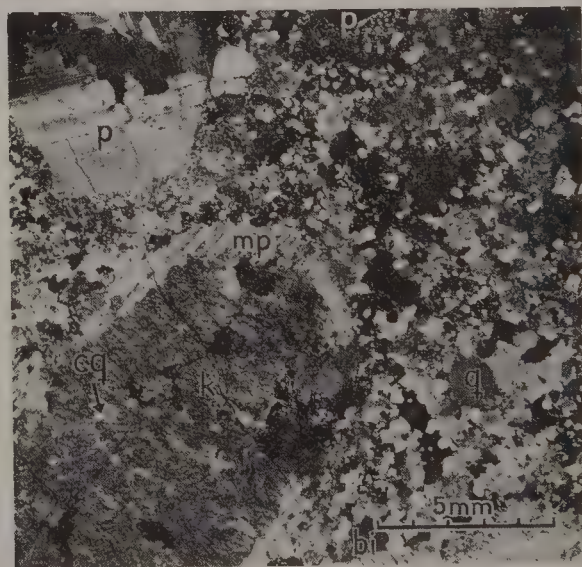


FIGURE 2. Photomicrograph of rapakivi texture, Wiborg massif, SE Finland: k, alkali feldspar ovoid; mp, mantling plagioclase; p, plagioclase; q, quartz; cq, concave quartz inclusions in ovoid alkali feldspar; b, biotite; p, spherical synneusis group of plagioclase crystals; XN.



FIGURE 3. Baltic shield rapakivi granite masses.

ley (1984). Examples of rapakivi texture in granitic terrane have been recorded from Australia, Austria, England, France, Norway, U.S.A., and U.S.S.R. and as far back as early Proterozoic times (Halliday et al., 1988).

Most granites with rapakivi texture are hypersolvus types, and the internal complexities of the phenocrysts indicate an involved crystallization history. As rapakivi textures vary from simple mantled phenocrysts to variants of orbicular structures, genetic theories of the two are linked (see Savolahti, 1956 for review). These include Sederholm's hypothesis for nucleation on impurities in the magma, Popoff's gravitative moment hypothesis involving concentric nucleation and accretion analogous to hailstone formation, Holmqvist's liquation hypothesis, Vogt's and Harker's hypothesis involving alternating resorption stages in a eutectic mixture, and Wahl's refusion hypothesis. Alternatives include, Wolff's hypothesis concerning hybrid magmatism, while Sederholm and Backlund consider metasomatic mechanisms to include rapakivi granite and various igneous and metamorphic rocks with orbicular structures. Tuttle and Bowen (1958), and Stewart and Roseboom (1962) consider experimental results and explain rapakivi texture as a special case of near eutectic, low-temperature recrystallization and re-equilibrium in the three-phase albite-anorthite-orthoclase region of the granite system.

ATSO VORMA

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Cross-references: *Batholith*; *Charnockite suite*; *Granite*; *Orbicular rocks*; *Phenocryst*.

REACTION PRINCIPLE AND REACTION SERIES

If crystals are in disequilibrium with the liquid when crystallization occurs in any silicate system, a reaction will occur between crystals and liquid, in which the liquid will attempt to redissolve the crystals so as to attain a state of equilibrium: this is the *reaction principle*.

Types of reaction relations

Continuous reaction series. During crystallization of any solid-solution series a reaction will occur between crystals and liquid. In Fig. 1 a liquid of composition A cools until it meets the liquidus, at which point crystals of composition Y are formed. When this happens, the liquid changes slightly in composition down the liquidus, and is then in disequilibrium with the original crystals at Y. A reaction then occurs between liquid and crystals, the crystals being resorbed into the liquid, whilst new crystals are precipitated at a slightly lower point on the solidus than Y. This process is a continuous one and, provided equilibrium conditions are maintained, the last of the liquid is used up at B when the crystals have a composition Z, i.e., all the crystals have the composition of the original liquid. Any solid-solution series may be termed a *continuous reaction series* during crystal-

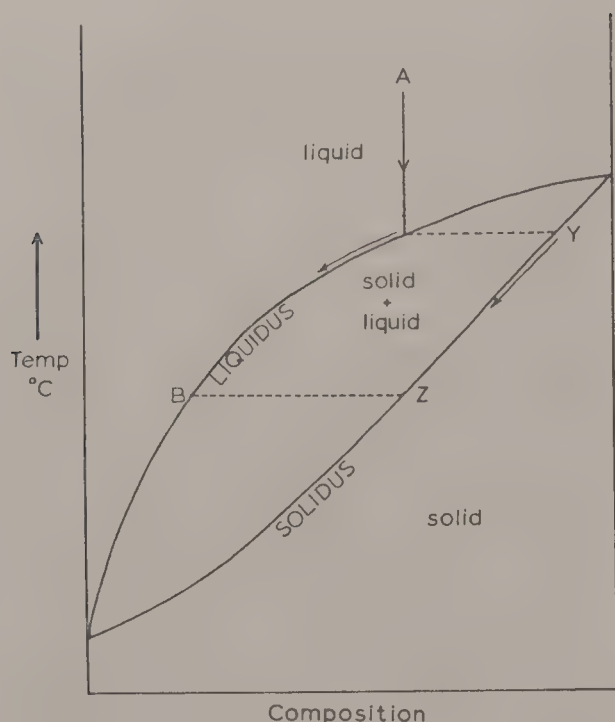


FIGURE 1. Crystallization of a solid-solution series.

lization. The olivines form a solid-solution series between forsterite (Mg_2SiO_4) and fayalite (Fe_2SiO_4). Another such series exists in the plagioclase feldspars between anorthite ($\text{CaAl}_2\text{Si}_2\text{O}_8$) and albite ($\text{NaAlSi}_3\text{O}_8$), but only at a pressure of 1 bar.

In any silicate system containing more than one solid-solution series, the process of continuous reaction between crystals and liquid will occur during crystallization in each of these series, even when the solid-solution series are incomplete.

Reaction pairs. Some binary systems contain compounds with incongruent melting points. In Fig. 2, A crystallizes first from the liquid of composition X and increases in amount as the temperature falls, until reaction point R is reached. At this temperature crystals of compound B form and the crystals of compound A react with the remaining liquid to produce more and more of B. Since the composition of the original liquid (X) lies between A and B, the final product consists wholly of crystals of A and B with no liquid remaining when the temperature reaches point R.

If the composition of the original crystallizing liquid lies between B and C, then crystallization continues to the eutectic at E and the final product consists of crystals of B and C.

Compounds A and B are termed a *reaction pair* and in the system $\text{SiO}_2\text{--Mg}_2\text{SiO}_4$, SiO_2 would represent C, Mg_2SiO_4 would represent A

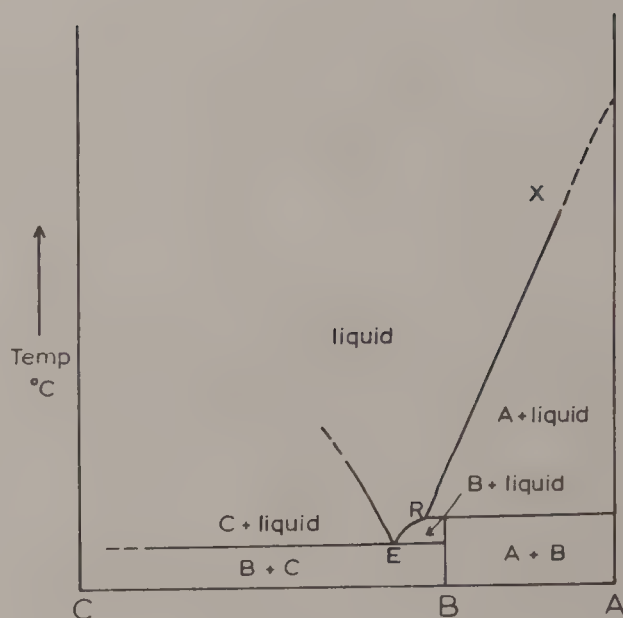


FIGURE 2. Crystallization of a binary system containing compounds with incongruent melting points.

and MgSiO_3 would represent B, a compound with an incongruent melting point.

Discontinuous reaction series. A binary system may contain several compounds with incongruent melting points. Figure 3 depicts the system $\text{K}_2\text{SiO}_3\text{--H}_2\text{O}$. K_2SiO_3 crystallizes first from the liquid of composition X and continues until point A, at which point a reaction occurs between crystals of K_2SiO_3 and liquid to produce crystals of $\text{K}_2\text{SiO}_3 \cdot \frac{1}{2}\text{H}_2\text{O}$. If fractiona-

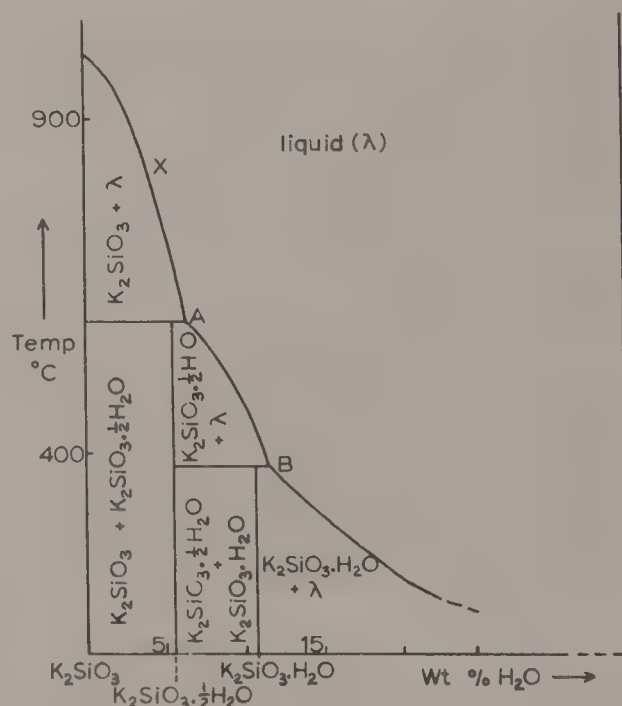


FIGURE 3. The system $K_2SiO_3-H_2O$. (After Morey and Fenner, 1917.)

tion occurs and the crystals and liquid are separated, further reaction between them occurs at point B when crystals of $K_2SiO_3 \cdot H_2O$ are formed. The compounds K_2SiO_3 , $K_2SiO_3 \cdot \frac{1}{2}H_2O$, $K_2SiO_3 \cdot H_2O$ constitute a *discontinuous reaction series*.

Whereas a continuous reaction series shows continuous compositional change during crystallization, a discontinuous reaction series shows compositional changes by means of a series of definite steps. There is no analogy between reaction series and eutectic crystallization: in reaction series the liquid does not solidify when a particular composition is attained, and there is a reaction relationship between crystals and liquid, not a subtraction relationship.

Reaction relations in continuous and discontinuous reaction series, and reaction pairs, are still retained when these become part of a more complex system. In reaction series, higher-temperature minerals appear before lower-temperature ones; thus a Ca-rich plagioclase occurs before a Na-rich plagioclase, and Mg-rich olivine before Fe-rich olivine.

Natural rock systems

The continuous reaction series in natural rock systems is shown by the plagioclase feldspars: where plagioclase crystals nucleating and growing do not continuously reequilibrate with the magma, zoned crystals develop with the cores more Ca-rich and the rims more Na-rich (Fig. 4). Continuous solid solution in the plagioclase feldspars is now known to occur at high temperatures only. The cooling rates of magmas control the kinetics of order-disorder in the feldspar lattice, determining the tetrahedral sites occupied by Al^{3+} and Si^{4+} cations. At low temperatures, phase relationships are complex

and feldspars rarely show complete adjustment to equilibrium, and perthites form with complete unmixing occurring at temperatures below $400^\circ C$ (Smith, 1974). The discontinuous reaction series is shown, generally, by the sequence (higher T) olivine-orthopyroxene-clinopyroxene-amphibole-mica (lower T) with variations in the compositions of the minerals dependent upon magmatic conditions (e.g., oxygen fugacity—see **Magmatic differentiation**). Where there is not continuous re-equilibration, *corona structure* (reaction rims) are formed, e.g., pyroxene around olivine and amphibole around pyroxene. Where early-formed minerals are separated from a residual magma (e.g., by sinking to the floor of the magma chamber and being covered by other crystals), rocks of contrasting composition are formed. An understanding of this process of fractional crystallization requires an awareness of the reaction principle and the concept of the reaction series. The same applies to the process of assimilation (contamination), where the relative positions in reaction series of minerals in the material being assimilated and of minerals crystallizing from the magma are a major control.

Related terms

Antipathetic—said of two or more minerals that are far apart from one another in a crystallization sequence and are not commonly found in association.

Broto crystal—applied to crystals (including xenocrysts) with corroded or embayed outlines.

Reaction border—a peripheral zone around one mineral that consists of another mineral formed by reaction of the magma with the earlier crystallized mineral.

Reaction rim—a reaction border seen in thin section (see **Reaction rim**).

Released mineral—a mineral formed during magmatic crystallization as a consequence of an earlier phase failing to react with the liquid; e.g., the failure of olivine to react with the liquid portion of the magma to form pyroxene may result in the enrichment of the liquid in SiO_2 which finally crystallizes as quartz (the released mineral).

Resorption—the partial or complete refusion or solution, by or in a magma, of previously formed crystals with which the magma is not in equilibrium owing particularly to changes in temperature and pressure (depth).

Resorption border—a corrosion border that represents partial resorption and recrystallization of previously crystallized minerals by magma.

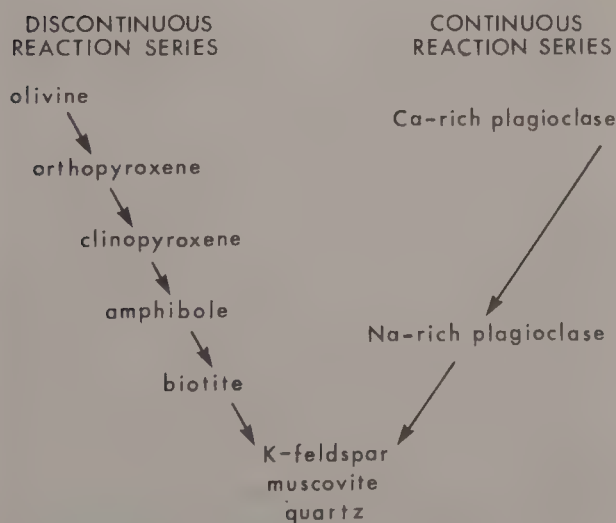


FIGURE 4. Discontinuous and continuous reaction series.

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Cross-references: *Assimilation*; *Cumulate textures*; *Magmatic differentiation*; *Reaction rim*; *Textures of igneous rocks*.

REACTION RATES IN METAMORPHISM

The rate of a chemical reaction is defined as the quantity of product formed in unit time. In this context, quantity may mean either weight, volume, or number of particles of product per unit volume. The reaction rate is dependent on the availability of reactant and the energy barrier involved in the bond redistribution processes necessary to complete the chemical transformation. Reaction rates are thus dependent on many variables, including (*inter alia*) temperature, pressure, concentration (availability and mobility) of participating elements (ionic species), deformation (which may exert a significant effect on diffusion rate), catalyst (both homogeneous and heterogeneous), and the ease of nucleation and growth of the product.

A major objective of metamorphic petrology is to provide a mechanistic description of the changes that occur during metamorphism. While the role of equilibrium in such transformations has been considered at length by the methods of thermodynamics, the significance of reaction rates has received much less attention. Nevertheless, there are strong reasons for considering that reaction rates, including the effect of nucleation and growth of a new phase, can influence both texture and, in some cases, the mineral composition of the assemblages formed. For example, if the relative rate of nucleation of a phase is high, this constituent of the rock may ultimately consist of a large number of small crystals. The kinetic controls of crystal size distribution have been reviewed by Spry (1969). The hypothesis that kinetic factors may determine which of two alternative minerals may form during metamorphism was advanced by Compton (1960) to explain the separate occurrence of cordierite and staurolite in different thermal aureoles. He postulated that cordierite was produced when rapid heating

overstepped the stability fields of slowly-forming, low-temperature minerals such as staurolite. However, where the staurolite-forming reaction was facilitated by water vapor, little overstepping occurred and staurolite formed first. If such behavior were common, the value of the usual thermodynamic interpretation of mineral assemblages would be seriously affected.

While this is not generally thought to be true, there are a large number of cases in which thermodynamic arguments seem inadequate to explain the observed assemblages. The possible significance of rate processes in some of these has been reviewed by Pitcher and Berger (1972). Perhaps the most common phenomenon that is difficult to understand on purely thermodynamic grounds is the survival of high-temperature assemblages throughout the long cooling histories during which they must traverse the fields of stability of low-temperature assemblages. This problem is discussed by Fyfe et al. (1958), who conclude that, in the hydrous systems common in metamorphism, the ratio of the forward to the reverse reaction might be as low as 2 and survival of the high-temperature assemblage would be unlikely. It follows, therefore, that either the forward reaction has been accelerated by undetermined kinetic factors or that the backward reaction has been inhibited. A reasonable assessment might be that the downgrade reaction is generally inhibited by the absence of water that acts as a catalyst. Deformation might, however, be expected to open up new diffusion channels and microcracks and it is well known that retrograde metamorphism, although a somewhat capricious phenomenon, tends to be associated with late-stage deformation.

Rates of metamorphic reactions may be considered through the well-developed methods of chemical kinetics, a field included in every text book of physical chemistry and the subject of numerous monographs. This approach depends on measurements of product yield vs. time data. This places a severe limitation on the application of such methods to metamorphic reactions, for which even time-temperature histories can only be inferred from indirect evidence based on assumptions regarding rates of heat flow, etc., and are at best semiquantitative. Reactions within or adjacent to a small dyke may be completed within a few days, whereas rocks may have been held at elevated temperatures for several millions of years during regional metamorphism. The time required to complete a given reaction may, however, be only a small proportion of the total time spent at the necessary temperature.

The study of reaction kinetics is carried out by measuring experimentally the progressive changes with time of the concentrations of reactants and products at known temperature. From the functions relating the variations of these concentrations with time, it is often possible to deduce the sequence of chemical steps involved in the overall transformation; this is the reaction mechanism. The most notable successes in the use of this approach have been in studies of homogeneous reactions (between gases and in solution), but an important branch has been concerned with reactions of solids. The participation of a solid reactant, possessing size, shape, and rigidity, introduces geometric parameters, since chemical changes almost invariably occur at a reaction interface that progressively advances into the solid reactant (Brown et al., 1980; Galwey, 1982). Information about the disposition and migration of such interfaces during formation of a phase in a metamorphic reaction may sometimes be inferred from examination of the products. The kinetic characteristics of such rate processes are not, however, so readily accessible, but a potentially powerful, and hitherto little used, approach is to compare, quantitatively if possible, the changes produced in experimental synthetic systems heated under pressure with those found in metamorphosed rocks. Assuming that there is some indication that comparable changes have occurred, the experimental system may be investigated further by measurements of the progressive changes in availability of appropriate constituents of the reactant mixture with time during isothermal reaction. The general equation applicable to the interpretation of reaction rates may be expressed

$$\frac{dP}{dt} = kf(C_A, C_B, \dots, P, \dots) \quad (1)$$

where dP/dt is the rate of product formation and C_A , C_B (etc.) are measures of the availability of the participants. Where reactions are autocatalytic, it is necessary to include the relevant function of concentration, area, etc., of products (P) in Eqn. (1). The precise form of this equation $\{f(\dots)\}$ will be determined by the reaction mechanism; if this relationship can be experimentally established for a particular system, it provides evidence for the nature of the rate-limiting step, and thus for the factors that exert greatest control on the reaction. A problem inherent in the study of reaction kinetics involving solids is the precise meaning that may be attached to the term concentration or availability of reactants. The magnitudes of such concentrations at the reaction site are not necessarily simple functions of the bulk concentration of the rock and may be controlled

by diffusion rates of various participants, solubilities in inter pore fluid, and the areas or types of lattice planes forming the surfaces of the reactants. The magnitude of P may, likewise, be dependent on several factors—distribution and shape of nuclei, ease of access of reactants, total area of product phase, etc.

While a general, overall form of Eqn. (1) that is applicable to every case cannot be meaningfully expressed, it is possible that in any particular reaction one or a small number of terms may be of overwhelming significance. Other necessary chemical processes are completed rapidly and the concentration terms do not enter the rate expression. A specific form of this equation thus provides evidence concerning the rate-determining step. Such simplification of the general expression is illustrated by the following example. If the rate of growth of a particular product phase, R , is controlled by the reaction of species Q , at concentration $[Q]$ in intergranular pore fluid, on the surface of the product area A_R , then the rate of the isothermal reaction could be expressed

$$\frac{dR}{dt} = k[Q]A_R.$$

However, if reaction depended (perhaps in a different temperature range) upon the rate of diffusion of an ion, S , to the reaction site, then the product yield, again under isothermal conditions, might be expressed

$$R = kt^{\frac{1}{2}}.$$

The rate constants (k) in these equations usually vary with temperature according to the Arrhenius equation (Brown et al., 1980)

$$k = Ae^{-E/RT}.$$

A , the pre-exponential term, is a measure of the frequency of occurrence of the event that is the precursor to possible reaction; it includes due allowance for the collision frequency between appropriate species, an entropy (steric) factor and a vibration frequency. E , the activation energy, is identified with the energy barrier to bond redistribution during reaction and is measured from the slope of a plot of $\log k$ against T^{-1} (T in kelvins). This interpretation of E may be applicable to a reaction in which overall rate is controlled by a single step, or to each individual step in the sequence where the reaction mechanism is complex.

In principle, detailed kinetic interpretation is capable of giving much information about the rate-limiting step of a reaction, but in practice it may be difficult to identify the particular chemical transformation that dominates the overall rate. Great care must be exercised in

assigning particular significance to a measured value of E if the reactants and products of a specific rate process have not been unambiguously identified (Galwey, 1982).

The principles of chemical kinetics have yet to be applied to more than a relatively few studies of metamorphic reactions. In addition to the above-mentioned work, examples of such studies include (1) the work of Brown et al. (1962), who used experimentally measured rates of the aragonite–calcite transformation to show that the temperature gradient during metamorphism and subsequent unloading of the Californian glaucophane schists could not have exceeded 10°C/km; (2) a review of the extensive literature relating to the kinetics and mechanisms of clay dehydration reactions by Brett et al. (1970); (3) the review by Lacy (1965) of factors involved in metamorphic reaction rates; (4) kinetic factors controlling the grain size distribution of garnet in various metamorphic rocks as discussed by Jones et al. (1972); (5) a hypothesis based on kinetic arguments advanced by Jones et al. (1972) to account for the variation in length/thickness ratios of biotite in metamorphic rocks; and (6) the study by Jones et al. (1975) of calcite recrystallization in the vicinity of a basalt dyke.

A. K. GALWEY
K. A. JONES

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Cross-references: *Metamorphic petrology—use of thermodynamics*; *Metamorphism*; *Nucleation of metamorphic minerals*; *Phase transformation*; *Retrograde metamorphism*.

REACTION RIM

Reaction rim is a genetic term for a border of secondary minerals formed at the margin of a primary grain in an igneous or metamorphic rock. Its use implies that the secondary minerals have formed as a result of reactions between the primary grain and the adjacent primary grain, a fluid, or a melt. Because a reaction rim is often seen to have partially replaced the primary grain, it is sometimes called a *corrosion mantle*. Nongenetic synonyms are *syntectic minerals* (minerals occurring at the contact of two grains) and *corona*, *atoll*, *moat* or *shell* where the rim entirely surrounds the primary grain. A rock which contains abundant coronas is called a *coronite* (Spry, 1969).

Petrography

A reaction rim is usually restricted to the grain boundary of a pre-existing single crystal, or to the margins of monomineralic mineral aggregates surrounded by minerals of contrasting composition. It can often be inferred that the rim mineral(s) have partially replaced the primary grains and they occasionally extend into them, usually along cracks. It may consist of one or more mineral species, distributed in one of a number of possible textural configurations. The simplest arrangement is a single-layer rim around a grain, such as the mantle of serpentine commonly found bordering olivine in peridotite (Fig. 1a). A simple reaction rim such as this may contain more than one mineral species combined in an intergrowth, e.g., in gabbroic or troctolitic igneous rocks olivine is commonly rimmed by a symplectitic intergrowth of pyroxene and spinel against primary plagioclase, and garnet in peridotite is often rimmed by a fibrous or symplectitic radial mass of pyroxene and spinel known as a *kelyphitic border* (Fig. 1b,c).

More complex reaction rims consist of two or more concentric layers, each of which may be either monomineralic or an intergrowth of a

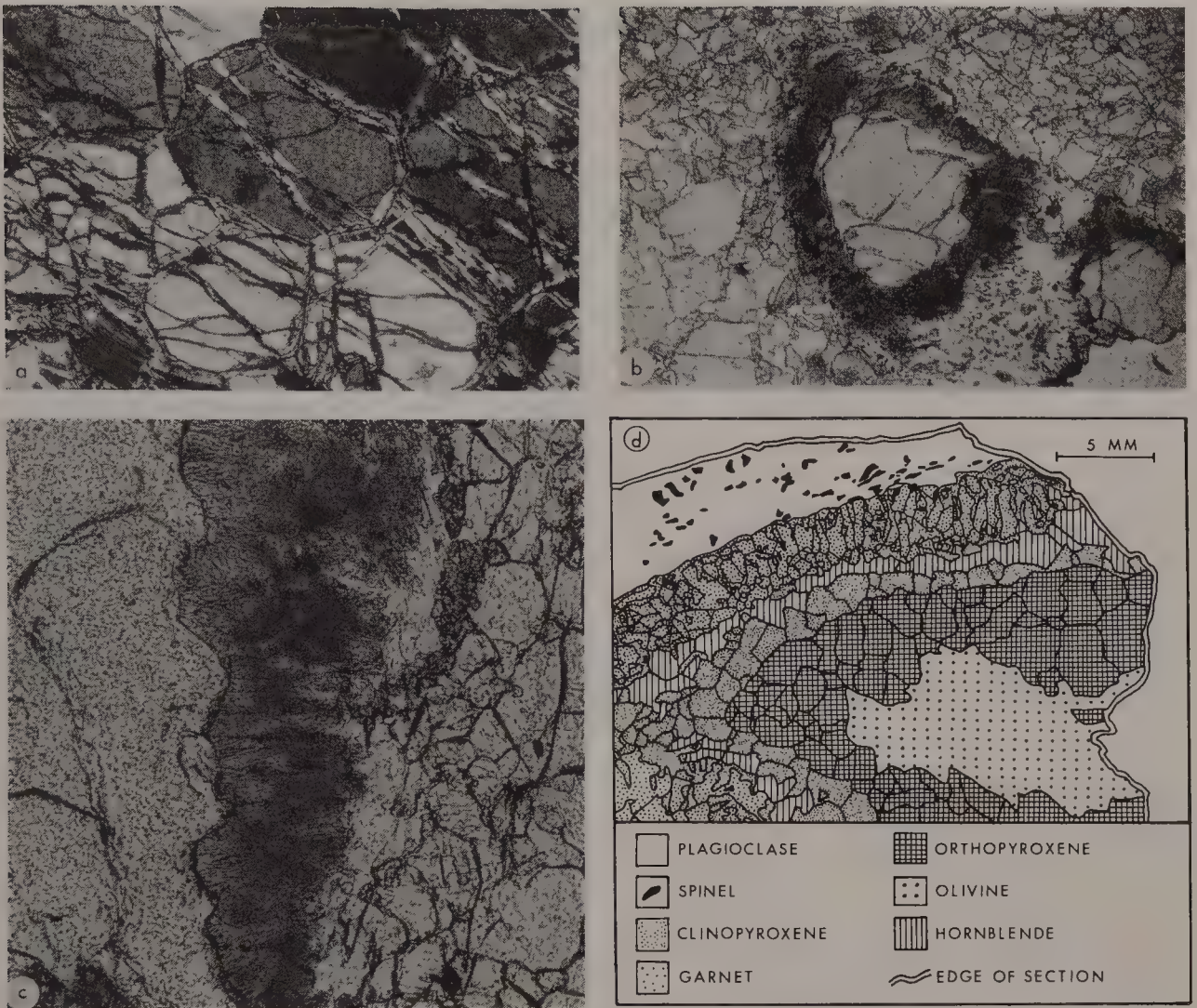


FIGURE 1. Examples of reaction rims.

(a) Olivine with reaction rims of serpentine along grain boundaries and cracks, resulting from the reaction of olivine with an aqueous intergranular fluid, XN, $\times 52$.

(b), (c) kelyphitic reaction rim around garnet in garnet peridotite; kelyphite is a symplectitic intergrowth of pyroxene and spinel (shown enlarged in (c)), resulting from the reaction of garnet with the surrounding olivine, PPL, $\times 22$, $\times 135$.

(d) Complex, multilayer reaction rim in meta-anorthosite, resulting from reactions between primary plagioclase and olivine, and subsequently between layers within the reaction rim itself. (After Griffin, 1971.)

number of minerals. The layers in these complex coronas may have formed contemporaneously or during two or more separate growth events, perhaps by reaction with each other as well as the primary grains. In any body of rock several stages of development of the coronas, frozen at different stages of growth, may be observed.

Some of the most commonly described complex reaction rims come from metamorphosed troctolitic igneous rocks. Griffin (1971, 1972) and Griffin and Heier (1973) have described examples of multi-layer, multi-stage coronas in meta-anorthositic rocks from Norway. Primary olivine is rimmed by several layers of secondary minerals against the adjacent plagioclase. In the most primitive type

of reaction rim olivine is surrounded successively by orthopyroxene, clinopyroxene, garnet, then a complex intergrowth of spinel, clinopyroxene and plagioclase against the primary plagioclase. In the more evolved type the garnet has overgrown the outer pyroxene + spinel layer, incorporating grains of the latter as inclusions (Fig. 1d). Progressive inward growth of orthopyroxene engulfs the primary olivine. Finally, a new layer of hornblende grows between the garnet and clinopyroxene layers.

In these examples three phases of reaction rim growth can be recognized: (1) pyroxene + spinel intergrowths and the orthopyroxene layer result from an early reaction between primary olivine and plagioclase, (2) reaction between the pyroxene + spinel intergrowth and the ortho-

pyroxene layer produces new layers of garnet and clinopyroxene, and (3) a reaction between the garnet and clinopyroxene layers in the presence of an aqueous fluid generates the hornblende layer.

If reaction rim formation proceeds far enough, some or all of the primary mineral species of the rock may be completely replaced. Due to the tendency for reaction rims to grow radially from the original grain boundaries, this will result in a new mineral association which mimetically replaces the primary mineralogy. Preservation of such *mimetic textures* demonstrates that these mineral transformations take place under conditions of little or no strain. It is not uncommon to find metamorphic rocks in low strain zones showing partial replacement of the primary minerals by reaction rims, adjacent to highly strained equivalent lithologies where the early minerals have been entirely replaced (e.g., Mørk, 1985).

Growth mechanisms

There are a number of possible causes for the formation of coronas (Spry, 1969): (1) primary magmatic accretion from a differentiating liquid, (2) reaction of early-formed crystals with a melt, (3) reaction in the solid state with an intergranular fluid such as deuteritic or juvenile water, and (4) reactions in the solid state between adjacent grains during regional or thermal metamorphism.

Coronas formed by primary magmatic accretion are not strictly reaction rims as they have not resulted from reactions at grain margins. However, it may be difficult to differentiate such rims from those resulting from the reaction of early-formed crystals with a melt. An example of the latter type may be the rim of hypersthene commonly found around olivine in tholeiitic basaltic rocks, where early-crystallized olivine reacted with a more evolved liquid than that in which it originally crystallized. However, some olivine phenocrysts in such rocks are rounded and have apparently been partially dissolved and resorbed into the melt. It would seem that growth of this type of reaction rim requires maintenance of a concentration gradient in the vicinity of the primary crystal during quenching (cf., Cox et al., 1979). Elimination of such a gradient by processes such as diffusion and mechanical mixing would inhibit development of a reaction rim and simply leave a resorbed margin.

The mechanism involving reaction with an intergranular fluid is most clearly demonstrated in essentially anhydrous rocks where reaction rims consist of hydrous minerals. A striking example of this type is the rims of serpentine

around olivine in peridotite (Fig. 1a). In the Norwegian coronites the layer of hornblende between garnet and clinopyroxene (Fig. 1d) grew at the expense of the latter two minerals during or after the introduction of an aqueous fluid into a previously "dry" rock.

A reaction rim resulting from growth in the solid state requires no input of externally-derived material, and is simply the result of reactions between adjacent minerals (under some circumstances an indigenous intergranular fluid may also be involved). The fact that such reactions take place between reactants which were previously in mutual equilibrium indicates that a change in externally imposed conditions (pressure, temperature) is responsible.

Several features of metamorphic reaction rims give clues to their mechanisms of formation. The growth of new minerals, as opposed to changes in the composition of existing minerals (compositional zoning), shows that the rim-forming reactions were discontinuous. The minerals comprising the rims replace the primary grains along their contacts, requiring dissolution of the primary grains and nucleation and precipitation of the new minerals in their place.

Metamorphic reaction rims usually form between domains of contrasting mineral composition (e.g., between plagioclase and mafic minerals in a gabbro). Once the rim has nucleated, further growth into one primary compositional domain requires transfer of material across it from the adjacent domain. For instance, during growth of garnet-orthopyroxene coronas between olivine and plagioclase, mafic components must be released by reaction at the olivine-orthopyroxene interface, pass through the orthopyroxene and garnet layers and be consumed by reaction at the garnet-plagioclase interface. Transport may be achieved by either intra-grain diffusion or intergrain (grain boundary) diffusion.

The occurrence of a reaction rim, as opposed to complete *pseudomorphism* (mimetic replacement) of the primary mineral by new metamorphic phases, indicates that the progress of reactions has been somehow retarded. Possible limiting factors may include a restriction in the availability of reactants and, related to this, a slow rate of diffusion through existing minerals of the various ions moving to the sites of precipitation. Minerals in a reaction rim are commonly chemically zoned, e.g., coronitic metadolerites from Sunnmøre, W Norway (Mørk, 1985) contain olivine with coronas of orthopyroxene and garnet against surrounding plagioclase. Garnet is enriched in Ca adjacent to plagioclase, while the plagioclase is depleted in Ca. Limited diffusion across the garnet layer

of Ca ions released from plagioclase may be responsible for their incomplete homogenization. Similarly, orthopyroxene is relatively enriched in Al adjacent to the garnet layer, suggesting only limited diffusion of Al from the plagioclase domain towards the olivine domain. Garnet may be a barrier to effective diffusion, so that growth of the reaction rim may itself create an impediment to further reaction. In the Sunnmøre examples, complete reaction produces simple eclogites (omphacite–garnet rocks) with unzoned minerals, indicating the operation of more effective diffusive homogenization, perhaps aided by deformation or transport in intergranular fluids (Mørk, 1985). A detailed quantitative treatment of the effects of diffusion during reaction rim formation is given by Mongkoltip and Ashworth (1983).

A metamorphic reaction rim commonly contains minerals which are not those expected in equilibrium assemblages for the particular bulk-rock composition and P – T conditions. The presence of orthopyroxene and sodic plagioclase in the Sunnmøre rocks is a good example of this, as such minerals are absent in compositionally equivalent non-coronitic eclogites nearby. The mineral associations in many reaction rims therefore appear to be a result of *local equilibrium* or *domain equilibrium*, due to the small scale of mass transfer in the absence of deformation or a fluid phase (cf., Koons et al., 1987).

S. J. CUTHBERT

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Cross-references: *Anorthosite*; *Atoll garnet*; *Eclogite*; *Gabbro and norite*; *Metamorphic mineral growth*; *Nucleation of metamorphic minerals*; *Peridotite*; *Reaction principle and reaction series*; *Textures of igneous rocks*.

REGIONAL METAMORPHISM

Metamorphic rocks that developed by recrystallization on a regional scale over large areas, hundreds of km² in extent, are the products of *regional metamorphism*. Three genetic types are distinguished: (1) regional dynamothermal metamorphism, (2) regional burial metamorphism, and (3) regional geothermal metamorphism. The first is recognized in eroded orogenic or mobile belts of various ages in which phyllites, schists, gneisses, migmatites, and granulites show the clear imprint of increased temperatures with the widespread development of penetrative deformation fabrics, consequent upon orogenic tectonism and often accompanied by the intrusion of magma. Regional burial metamorphism may be associated with ocean-trench environments and subduction processes or it may bear no direct relationship to orogenesis or to magmatic intrusions. Deep burial of sedimentary volcanic rocks during the development of geosynclines or continental basins may eventually give rise to limited recrystallization in response to the prevailing geothermal gradients, without penetrative deformation. When burial is accompanied by high load pressures, total reconstitution of the buried rocks may be accomplished. The main expression of regional geothermal metamorphism is recognized in mid-oceanic ridge (ocean floor) situations (Miyashiro et al., 1971) but may also be expected in aulacogens and rift zones where magmatism and high thermal gradients bear witness to increased thermal input without penetrative deformation. Regional increase in temperature and pressure resulting from the depression of rocks to great depths or through tectonic thickening, clearly transfers the rocks from the original crustal setting in which they were deposited or emplaced and promotes the development of new rock fabrics and minerals assemblages in equilibrium with their new physical and chemical conditions. These products of regional metamorphism therefore represent a frozen record of the tectonothermal history of the Earth's crust since early Archaean time.

Controls, processes and causes

Temperature, hydrostatic pressure, and shearing stress, together with the chemical activity of percolating pore fluids, are the major

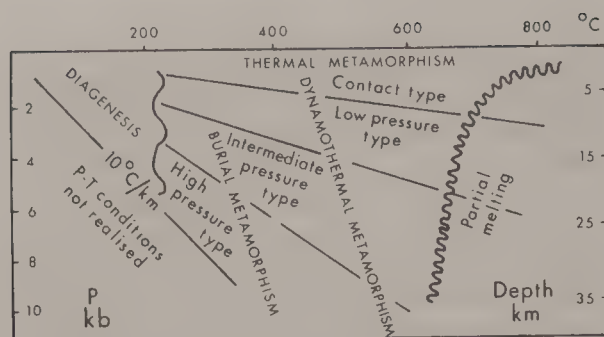


FIGURE 1. Simplified pressure-temperature diagram for different types of metamorphism. (After Winkler, 1967.)

physical variables governing the process of regional metamorphism. The temperature and pressure conditions vary between the two extremes of diagenesis and partial melting (anatexis) and can be represented graphically on a pressure-temperature diagram (Fig. 1). Temperature is the principal control and in accordance with the prevailing geothermal gradients may range from 300 to 900°C while complementary hydrostatic (load) pressure values are less certain but may reach 12 kb. Directed shearing stresses determine the penetrative deformation fabrics and structures of the rock, and thereby facilitate the passage of pore fluids that catalyse the chemical reactions leading to new stable mineral assemblages. Partial pressures of H_2O , CO_2 , O_2 and other fluids may give rise to overpressures and hydraulic fracturing that locally influence the equilibrium temperatures of the hydration and dehydration reactions that characterize the metamorphic process. The establishment of fluid pressure gradients may initiate diffusion, thereby assisting the processes of metasomatic alteration and metamorphic differentiation.

Regional metamorphism is developed in a broad spectrum of crustal conditions as the result of influxes of heat and abnormal but variable regional geothermal gradients that may be generated by

- (1) radioactive disintegration;
- (2) enhanced convective heat flow from the mantle;
- (3) subduction and lithospheric plate motion;
- (4) recrystallization at increased temperature due to deep burial;
- (5) convective heat flow due to a regional invasion of country rocks by granitic and/or basic magma;
- (6) regional metasomatism and/or invasion by volatiles;
- (7) deformation under tangentially direct pressure; and

- (8) frictional heat along major shear zones promoting the rise of granite into the crust.

Thus, not only do burial or deformation alone bring about a change, but accession of heat to the crust is essential to promote metamorphic reactions, the ultimate source of heat being sought in thermal foci originating in the mantle.

Because heat and deformation are the results of different processes, they are essentially independent variables in any scheme of regional metamorphism. The two-fold response of rocks to these factors is expressed in the time relationships between the development of new mineral parageneses and rock fabrics consequent upon the varying rates at which they operate.

Recognition of progressive metamorphism

Metamorphism is by no means a general feature of all strongly deformed rocks. Certain shallow orogenic belts may lack metamorphic rocks; in others, recrystallization has been accomplished at low temperatures, whereby sediments have suffered limited prograde change and intercalated volcanic rocks have undergone only progressive hydration. In most regional metamorphic regimes, however, a systematic increase in metamorphic grade is seen in broad progressive patterns, with the rocks displaying tectonic styles appropriate to the various geodynamic environments found within the orogenic zone. In rocks of given composition, the successive stages of metamorphism can be followed as diagnostic index minerals make their first appearance. A succession of typomorphic minerals, as metamorphic grade increases, can be distinguished on a regional scale in some classical areas, such as in the Caledonides of Scotland, where the Barrovian prograde sequence of chlorite → biotite → almandine garnet → staurolite → kyanite → sillimanite is typical in pelitic Dalradian rocks. These index minerals characterize metamorphic zones, their boundaries representing the intersection of isothermal surfaces with the present-day topography (Fig. 2). In rocks of highest grades, partial melting at elevated temperatures has resulted in the development of migmatites and syntectonic granitoids. Other types of metamorphic zones have also been recognized as in the Hercynian belt of the central Pyrenees (chlorite → biotite → staurolite + andalusite + cordierite → andalusite → cordierite → sillimanite → cordierite) and in the high-pressure regions of the Franciscan terrane of the western U.S.A. (chlorite-stilpnomelane → pumpellyite → glaucophane → lawsonite-jadeite → pyralspite garnet → piemontite).

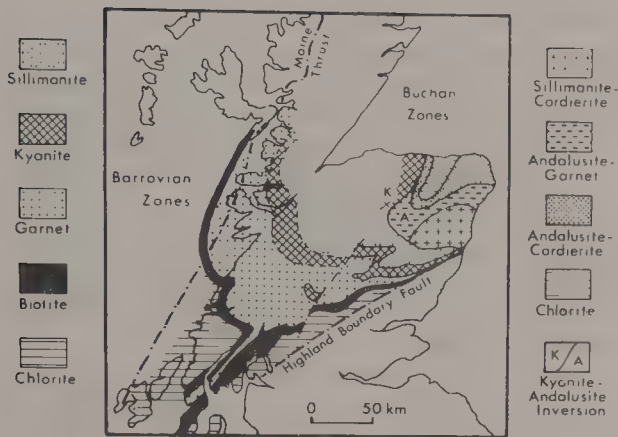


FIGURE 2. Metamorphic zones in the Scottish Caledonides. (After Winkler, 1967.)

Deformation and metamorphism are independent variables in the orogenic scheme of the Michigan region, U.S.A., where heat from subjacent bodies was responsible for metamorphic patterns that show no clear relationship to regional structures or to granitoid intrusions of the same orogenic cycle (James, 1955).

Despite the wide compositional range of metamorphic rocks, mineralogically they are easily classified in terms of sets of simple regular mineral assemblages dependent upon original chemical compositions. Hence, rocks of any composition that have been metamorphosed within certain broad limits of temperature and pressure belong to a single metamorphic facies as first defined by P. Eskola and subsequently modified by others (e.g., Turner, 1981). Thus, the different trends of regional metamorphism may be described with respect mainly to temperature by identifying the various metamorphic facies in one region (see **Metamorphic facies**).

Winkler also recognized that an assemblage of minerals is petrologically more significant than the occurrence of one mineral, which is the essence of the Barrovian scheme. He subsequently suggested, however, the abolition of the term facies and recommended that the entire P - T range of metamorphic conditions be divided into four divisions of metamorphic grade—very low, low, medium, and high grades—the boundaries between them being marked by significant changes in mineral assemblages or isoreaction grads (Winkler, 1970). Within each division, the isoreaction grads in common rocks constitute a metamorphic grid, thereby serving to classify the whole field of metamorphic conditions. Boundaries between Winkler's divisions reflect changes that have been recognized to take place at the beginning of the greenschist and amphibolite facies, respectively.

The modified concept of depth zones as formerly proposed by Grubenmann and Niggli (1924) classified metamorphic rocks into three categories according to physical characteristics in order of increasing depth of metamorphism. Metamorphism at shallow levels of the epizone was assumed to be characterized by low temperature, low pressure, and strong deformation; that at the katazone or deepest zone by high temperature, high pressure, and weak deformation, and there was assumed a mesozone of intermediate characteristics. Although these terms are still loosely applied, it can now be appreciated, through modern studies, that a simple and direct relationship between metamorphic conditions and depth is no longer valid. Instead tectonic styles and rock fabrics show a marked conformity with appropriate metamorphic grade in normal progressive metamorphic zonation and polymetamorphic relationships. Characteristically, low metamorphic grades in orogenic zones are accompanied by schistose fabrics and brittle failure giving way to increasing plasticity in gneissic rocks and the development of migmatites as metamorphic grade increases. At the very highest grades of the granulite facies, granoblastic and flaser textures, mylonitic fabrics and flow-folds are common. Such systematic changes in the mineralogy and/or mineral assemblages and rock fabric are dependent upon the original composition of the rocks involved as well as fluctuations in both regional temperature and/or pressure gradients caused by thermal disturbance in the lower crust or mantle. Thus in many orogenic belts the thermal structures appear to be relatively simple.

Pressure factor and trends

Variation in the metamorphic style of different orogenic belts is a function not only of temperature but also of pressure. The term *facies series* was introduced by Miyashiro (1961), to denote the series of metamorphic facies produced in a metamorphic region under the influence of a specific geothermal gradient. At least five major types were recognized, of which the Barrovian sequence, previously considered to be the "normal" result of regional metamorphism, is but one example (Fig. 1). Hietanen (1967) extended the known range of metamorphic categories by recognizing at least nine common types that are characteristic for particular temperature-pressure regimes (see **Metamorphic facies series**).

The regions in which individual trends have developed may range in extent from entire fold belts or large parts of mountain chains distinguished by one metamorphic trend, to

small tectonic units of orogenic belts predominantly affected by different trends. For example, in NE Scotland the area of Buchan-type metamorphism is marked by typomorphic andalusite, sillimanite, and cordierite developed late in the tectonic history of the more extensive adjacent Barrovian-type metamorphism (Fig. 2).

Duality in the pressure regime in which relatively higher-pressure facies series are marginal to a central area developed under lower-pressure, higher-temperature conditions is recorded in many belts of all ages. These polymetamorphic features are not necessarily coeval but rather the higher-pressure regime prevails, at least in part, before the establishment of lower-pressure conditions. Such duality is the symmetrical expression of the total metamorphic regime within a single belt in which the central high-temperature core with its attendant granitoids is the response to the greater heat flow along the mobile belt axis.

Collectively, the trend of metamorphic facies series within an orogenic belt may show a progressive development that cuts across the traditional facies series as defined by Miyashiro (1961). Such a trend is noted for the rocks of the Grenville Province of Canada and has been termed a transcurive facies series by Chesworth (1972). Although, for the Grenville Province, the transcurive series is apparently discontinuous in place and made up of fragments of the conventional facies series separated by faults, the evidence in southern Africa suggests that transcurive trends, where recognized, overstep or are probably continuous both in space and time. These trends are in marked contrast to metamorphic patterns within the circum-Pacific paired belt regime where two contrasting belts with their own temperature-pressure regimes and distinctive igneous associations are juxtaposed parallel to one another (Fig. 3). One belt is characterized by high-pressure-low-temperature metamorphism

(Franciscan or Alpine type), often contains glaucophane schist (blueschist) and abundant basic and ultrabasic rocks, and occurs along the oceanic margin. The other belt, situated on the continental side, is characterized by relatively low-pressure-high-temperature metamorphism with abundant syntectonic and late-tectonic granitoid rocks (Abukuma type). Together these paired metamorphic belts appear to relate to a subduction environment (see **Paired metamorphic belts**).

Thus, metamorphic complexity may vary from simple progression of variable intensity to metamorphic duality, paired belts, and more frequently polymetamorphism that is a consequence of (1) polyphase metamorphism within a single event, (2) polycyclic metamorphism (polymetamorphism) in successive events and (3) retrogressive metamorphism resulting from unloading and/or declining temperatures.

Yet a further comparison between metamorphic trends of older cratonic regions and those of the younger orogens suggests an evolutionary sequence in which the temperature/pressure ratio decreases with time (Saggerson and Turner, 1972).

Metamorphic regime

Within the concept of plate tectonics several orogenic environments of regional dynamothermal metamorphism can be recognized: ocean floor-island arc collision, island arc-island arc collision, ocean-continent collision, island arc-continent collision, continent-continent collision, mid-ocean ridge-continent collision, aulacogens, intracontinental mobile belts, Archaean cratonic nuclei, and greenstone belt association. Regional burial metamorphism, however, develops within supracrustal rocks of deep continental troughs often associated with and adjacent to orogenic zones, in geosynclinal successions that were never involved in orogeny, and in subducted plates.

Each mobile belt presents a unique situation, but some common factors can be noted. Dietz (1972) has emphasized that it is primarily along the trailing edge of a drifting continental plate that the great geosynclinal prisms are deposited. With the development of an oceanic trench and subduction zone, the continental margin and trench would eventually collide, causing collapse of the eugeosyncline and the tight folding and dynamic metamorphism of the sedimentary pile (Fig. 4), while the miogeosynclinal successions, protected by the stable continental slab, would be only mildly folded and metamorphosed to low grades under the main influence of regional burial metamorphism. Coleman (1972) maintains that the presence of blueschists

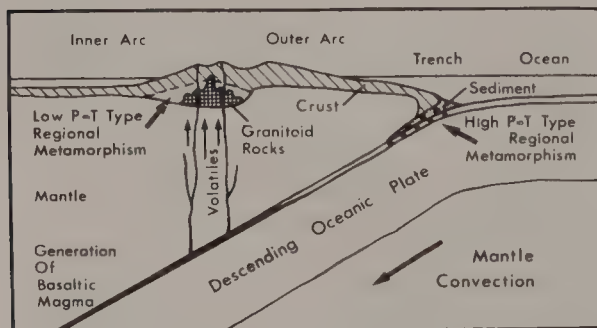


FIGURE 3. Paired metamorphic belts. (After Miyashiro, 1973.)

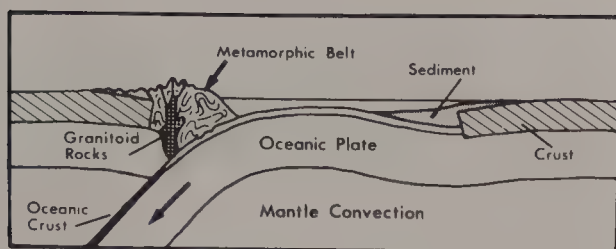


FIGURE 4. Metamorphic regime in an orogenic belt of Atlantic margin type. (After Dietz, 1972.)

in narrow belts can be related in space and time with geosutures developed on the former edges of interacting lithospheric plates. The low thermal gradients resulting from rapid burial to great depths of a cold, dense, descending prism of trench sediments and ocean floor combined with the unusually high pressures and tectonic thickening developed in the compressional zones, where subduction and obduction are active, provide the optimum conditions for blueschist formation. Associated ophiolites represent former oceanic crust caught within a mélangé zone. The high pressures are a consequence of the subduction environment developed within a cold, descending prism of trench sediments and ocean floor carried to great depth or under obducted oceanic crust. Low-pressure metamorphism is developed within the geosynclinal prism situated on the continental side of the subducted zone.

Frequently, a series of geosynclinal furrows are welded together. In such furrows three main tectonic units may be recognized, a foreland facet including any miogeosynclinal development and a eugeosynclinal facet separated by a miogeanticlinal rise (Fig. 5). Each exhibits its own contrasting metamorphic characteristics yet all relating to the total temperature–

pressure regime. Within the cratonic regions of the continents, the present distribution of regional metamorphism indicates a polymetamorphosed sialic foundation to the former orogenic belts that developed between continental plates (Kröner, 1977).

The geodynamic history of any orogenic belt involves the interplay of tectonic adjustment and a thermal influx, both originating from the same activating processes, but reaching peaks at different times and in different places. The influence of pressure and/or temperature will necessarily create variety in the metamorphic and structural response according to position (tectonic level) within the orogenic belt. Within the foreland facet, where magmatism is generally lacking and folding is minimal, the very low-grade metamorphism in miogeosynclinal basins can be largely attributed to burial in response to the regional geothermal gradient. Where the marginal tectonic effects of the orogenic zone can be demonstrated in these rocks, they are usually accompanied by metamorphic upgrading in response to additional heat flow derived from the adjacent orogenic zone.

In the transition area between the foreland and eugeosynclinal areas where major tectonic movements such as overthrusting and shearing are often evident, the associated metamorphic element is often represented by the highest-pressure assemblages pertaining to the earliest stages of development of the belt. Progressive regional metamorphism increases markedly away from the cratonic foreland into the mobile zone, the main development of a facies series thus occurring within the eugeosynclinal facet.

The areas of highest grade are also areas of potential granite development and an abatement of the regional pressure factor will frequently give rise to low-pressure metamorphism, porphyroblastesis, widespread granitization, and, in time, the eventual mobilization and late tectonic intrusion of granitoid rocks and pegmatites. Superposed contact metamorphic aureoles about intrusive granitoid rocks are a feature of this stage. The tectonic framework of these low-pressure areas is generally more diffuse than the predominantly linear aspect of the higher-pressure regions of the orogenic belts.

Geodynamically it means that total rock pressures are significant during the early stages of the development of a mobile belt, but as rising geoisotherms invade the crust, the response in the rocks with time, as denoted by new mineral assemblages and changing tectonic styles, is governed largely by temperature, thereby reducing the effects of the prevailing pressure gradient.

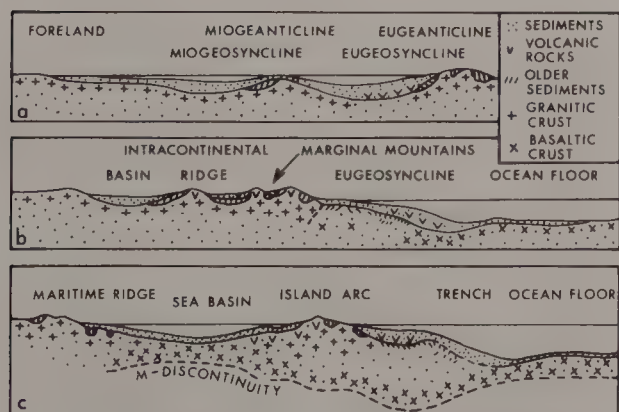


FIGURE 5. Tectonic framework of orogenic belts: (a) Europe, (b) Mesozoic examples in E Asia, (c) Cenozoic examples on the western side of the Pacific Ocean. (After Wyllie, 1971.)

Timing

Within the life span of an orogenic belt, the processes of orogenesis involve a succession of phases of deformation, recrystallization, and granite emplacement representing episodes of a polyphase event. The duration of such episodes, and therefore the life of a metamorphic belt, can vary considerably. In some mountain chains, as the Alps, it may be as short as 30 Ma. For Precambrian belts periods of as much as 900 Ma have been reported from some, while for others it is comparable with that of Phanerozoic belts: in the early Proterozoic Svecofennides of S Finland, Hopgood et al. (1983) demonstrated an extensive metamorphic, migmatitic and structural history to have taken place in <20 Ma. This regional metamorphic episode, like others attributable to collision tectonics, is apparently of relatively short duration and is in marked contrast to the metamorphic history of intracratonic orogenic belts that have developed over much longer periods of time.

Most orogenic belts are polyphase. In some, overprinting is detected merely through thermal resetting of radiometric dates but in the majority of cases widespread metamorphism, intrusion, and tectonism have occurred during successive phases. Regional low-grade retrogressive phases may accompany the final stages of a long and complex period of metamorphism, tectonism, and uplift but it may equally be the response in cratonic rocks to metamorphic reworking within the margins of peripheral orogenic belts, or is recorded where transition from craton to tectogene is marked by major faults or phyllonitic shear zones. A characteristic sequence is one in which primary folding with development of nappes is followed by regional metamorphism that may reach a peak during a subsequent phase of less intense refolding. Plutonic activity including emplacement of both basic, and in particular, acidic rocks occurs intermittently throughout the duration of any mobile belt.

The present distribution of remnant cratons and mobile belts indicates a prolonged history of metamorphism and tectonism in recurring cycles throughout the development of the cratonic crust. This progressive growth in the development of many ancient metamorphic terranes is evidenced by polycyclic metamorphism, development of granulite facies rocks, the common relationship between infrastructure and suprastructure, remnants of old rocks in younger metamorphic terranes, superposition and cross-cutting relationships of mobile belts of different ages, remobilized sialic crust, mantled gneiss domes, changes in isotopic ratios, and concentration of elements of economic

importance.

Regional metamorphism relates intrinsically to mobile belts, and their history can be unraveled by studying time relationships between burial, deformation, metamorphism, and their association with enhanced thermal input from intrusions and subsequent uplift with erosion. Pre-, syn-, late-, and post-tectonic phases in the life span of an orogenic event can rarely be divorced from the varied regional metamorphic response in the different crustal environments. Although it is possible to relate regional metamorphism to modern global tectonic situations, it is often more difficult to apply the same techniques to older belts, particularly those of Precambrian age. Whether this is the result of inherent differences in mobile belts with time, or difficulty of application of the techniques to much deeper levels of the belts, is a matter for further study.

E. P. SAGGERSON

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Cross-references: *Burial metamorphism*; *Island arcs—petrology*; *Metamorphic facies*; *Metamorphic facies series*; *Metamorphism—controls*; *Ocean-floor metamorphism*; *Paired metamorphic belts*; *Retrograde metamorphism*.

RETROGRADE METAMORPHISM

Retrograde metamorphism (diaphthoresis, retrogressive metamorphism) is the mineralogical adjustment of relatively high-grade metamorphic rocks to temperatures lower than those of their initial metamorphism. A *diaphthorite* is a metamorphic rock in which minerals characteristic of a lower metamorphic grade developed at the expense of minerals formed at a higher metamorphic grade.

The term *diaphthoresis* (from the Greek “to degrade”) was first applied by Becke (1909) to a phyllonite containing relict high-grade metamorphic minerals and relict gneissic layering in an otherwise low-grade mineral assemblage. The approximately equivalent term *retrograde metamorphism* is now commonly applied to metamorphic rocks that display any degree of replacement of a high-temperature mineral or mineral assemblage by one stable at a lower temperature. Where the process has not been carried to completion, the disequilibrium mineral assemblage will give evidence of the polymetamorphic history. In other cases, adjustment to the new conditions may have been complete and the retrograde nature of the rock can be deduced only if the low-temperature minerals pseudomorphically preserve the forms of the minerals of the higher-grade rock (e.g., chlorite after garnet—Fig. 1).

The term retrograde literally refers to a return to a prior or inferior state and consequently should be restricted to rocks for which the high-temperature past was truly a metamorphic one (Turner, 1948). The term diaphthoresis does not carry the implication of return to a prior state and thus could be used also for the metamorphism of igneous rocks at temperatures lower than those of their initial crystallization. Winkler (1967) has suggested

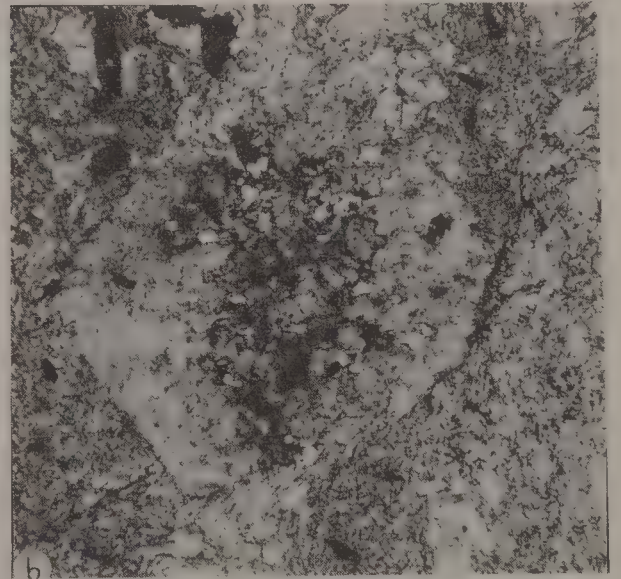
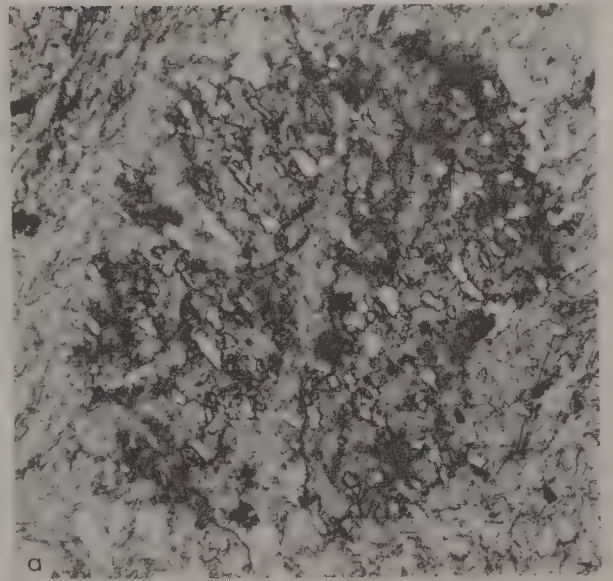


FIGURE 1. Selective replacement of a poikiloblast of garnet (dark) by chlorite (a) leading towards the development of a pseudomorph of chlorite after garnet (b); garnet-mica schist, Ballachulish, Scotland, PPL, $\times 17$.

that burial of volcanic rocks intercalated with sediments may result in retrograde metamorphism of the volcanic rocks at the same time as the sedimentary rocks are being progressively metamorphosed. This is true in the broader literal meaning of the term diaphthoresis but not in the more restricted meaning of the English term. There is no single English equivalent for that part of diaphthoresis that deals with metamorphism of igneous rocks, although some cases are covered by the terms autometamorphism and hydrothermal alteration.

Retrograde metamorphic reactions

Many authors have cited examples of one or more low-temperature minerals replacing original minerals of a higher grade as examples of retrograde metamorphism: Table 1 lists common examples. No attempt has been made to write balanced reactions for these replacements, since the actual course of the reaction is generally unknown.

Both polymorphic transitions and exsolution reactions occur in response to lowered temperatures. Harker (1950) has referred to exsolution perthite as resulting from retrograde metamorphism, but such reactions are better considered as separate phenomena and have not been listed in Table 1.

TABLE 1. Retrograde metamorphic reaction pairs

Reactant(s)	Product(s)
Actinolite	Clinocllore or antigorite
Andalusite	Sericite or muscovite + quartz
Andradite	Epidote
Biotite	Chlorite + magnetite
Cordierite	Chlorite + sericite or muscovite
Corundum	Diaspore or margarite
Diopside	Tremolite or actinolite
Forsterite	Serpentine
Garnet (almandine)	Chlorite
Grossular	Idocrase, zoisite or clinozoisite
Hornblende	Biotite, Chlorite or Serpentine
Ilmenite	Leucoxene
K-feldspar	Sericite
Kyanite	Sericite or muscovite + quartz
Periclase	Brucite
Plagioclase	Zoisite or clinozoisite + albite
Sillimanite + biotite	Staurolite + muscovite
Sillimanite	Sericite or muscovite
Staurolite	Chloritoid or muscovite + quartz
Wollastonite	Calcite

Controlling factors

The frequent failure of high-grade metamorphic rocks to experience more than incipient retrograde metamorphism during slow cooling has been explained in terms of three factors that appear to be most important in controlling the rate of retrograde metamorphic reactions (cf., Vernon, 1975; Turner, 1981).

Temperature. For chemical reactions taking place at temperatures far removed from the equilibrium temperature, the Arrhenius equation shows that the rate of reaction will be greatly increased by increasing temperatures. Near the equilibrium temperaure, however, the reaction rate is also determined by the free-energy change of the reaction. The free-energy change and the reaction rate are both zero at the equilibrium temperature. As

an equilibrium boundary is passed in the prograde direction, both the increasing temperature and the increasing free-energy change of the reaction will tend to increase the reaction rate. On crossing the same boundary in the retrograde direction, the increasing free-energy change will be working in the opposite direction to the falling temperature; the former will increase the rate of reaction while the latter will decrease it. As a consequence, prograde reactions will usually proceed at higher rates than will the same reactions in the retrograde direction and will be more likely to be carried to completion.

Absence of a fluid phase. As pointed out by Schwartz and Todd (1941), nearly all retrograde metamorphic reactions result in an increase in the amount of water (or occasionally CO₂) contained in the mineral assemblage. An aqueous fluid phase also serves as a medium for diffusion, increasing the rate of reactions in which it is not produced or consumed. If the original pore fluids and those released by chemical reactions are driven out of a rock during prograde metamorphism, so that most of the water that remains is chemically bound in the minerals of the high-temperature assemblage, and if the rock becomes essentially impermeable during the initial metamorphism, retrograde metamorphic reactions will be hindered or prevented. However, Turner (1948) contends that water is present in excess during metamorphism and is a mobile component capable of freely entering or leaving the reacting system. Hyndman (1972) has suggested that hydration of minerals of high-temperature assemblages during the early stages of retrograde metamorphism results in their expansion. The volumetric increase of mineral constituents would then seal off the pores and thus prevent the continuation of the retrograde metamorphic reactions.

Penetrative deformation. Dachille and Roy (1960) demonstrated that metamorphic reactions being run at low temperatures in a simple squeezer device can be greatly accelerated by applying a rotational shearing motion to one of the anvils. This shearing may increase the reaction rate by a number of different mechanisms, such as increasing the surface area and strain energy of the reacting mineral grains, exposing fresh surfaces on the grains to reaction, and increasing the number of nuclei present. If the thermal peak of metamorphism is reached concurrently with or following the final phase of deformation, this "catalytic" effect of shearing will be limited or absent during the waning stages and the high-grade mineral assemblage is more likely to be preserved. Knopf (1931) has noted that many of the rocks

that have experienced extensive retrograde metamorphism, including Becke's original "diaphthorite," are *phyllonites*. These rocks have been metamorphosed in regions that were being highly deformed. In such cases the deformation may have been directly responsible for the retrograde metamorphism, or movements associated with the deformation may have led to the aqueous fluids having access to the rocks.

D. W. CHIPMAN

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Cross-references: *Greenschist facies*; *Hydrothermal processes*; *Metamorphic Reactions*; *Phase transformation*; *Phyllite*; *Polyphase metamorphic mineral growth*; *Reaction rates in metamorphism*; *Regional metamorphism*.

RHEOMORPHISM

The word "rheomorphose" (Greek *rhein*, to flow) was introduced by Backlund (1937) and *rheomorphism* was defined as the whole process by which a rock is converted partly or entirely into liquid with the entry of more or less new material that has diffused in. Mehnert (1968) defines rheomorphism as the "process of at least partial mobilization involving addition and removal of diffusing elements such that a rock becomes mechanically mobile" (p. 356), while Reynolds (1947) indicates that rheomorphic "means 'flow form', and signifies that some pre-existing rock, whether initially sedimentary,

plutonic, metamorphic, or volcanic, has acquired from the solid state sufficient fluidity to become eruptive as a result of chemical changes and rise of temperature" (p. 212).

The qualification relating to the addition of new material into the rock was added because Backlund (and others) were working in metamorphic terranes where migmatization and granitization are extensive, and thus where granitization and rheomorphism tend to go hand in hand. However, the term has been broadened by many workers to include mobilization of rocks in environments other than migmatitic ones, for instance, mobilization of rocks by fluidization processes and the mobilization of hornfelses in high-temperature contact metamorphic aureoles.

Because rocks that may never have been connected with a fluid phase can show evidence of "flow forms," such as some rocks deformed by high pressure, and salt within salt domes, Mehnert (1968) suggested that the term "rheomorphism" be used only in its purely literal sense, i.e., as an indication of the pronounced increase in mechanical mobility of otherwise solid rocks. The term was specifically selected to exclude any hypothetical assumption as to the state of the material during emplacement and not as a synonym for "melting" (Holmes, 1965). However, the definition of Backlund (1937) includes the word *Verflüssigung*, which means liquefaction.

The terms "mobilized replacement dyke" and "rheomorphic dyke" were suggested by Goodspeed (1952) for dykes where flowage of metasomatized material is clearly indicated and subsequently (1953) he described a series of rocks in a hornfels–granodiorite contact zone to which the term "rheomorphic breccia" is given and tabulates differences between these breccias and breccias of igneous plutonic and replacement origin. Earlier (1939) he had introduced the term "neomagma" for the material formed by the process of rheomorphism where any solid rock is made to flow as the result of chemical changes and an increase of temperature.

Relatively large-scale rheomorphism in areas of regional metamorphism has been described from a number of localities. In Donegal, for instance, dome-like intrusions of granite are referred to by Pitcher (1970) as "rheomorphic domes" and ascribed to the mobilization and migration of crustal rocks. Similarly, in the Trois Seigneurs Massif of the French Pyrenees, masses of quartz diorite are considered to have resulted from the rheomorphism of sillimanite gneisses (Allaart, 1959).

A. R. WOOLLEY

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Cross-references: *Diapirism*; *Fluidization*; *Intrusion*; *Migmatite*.

RHYOLITE

Rhyolite is a fine-grained or glassy igneous rock chemically and mineralogically the volcanic equivalent of granite. The name was originally applied to such rocks by von Richthofen (1860, see Johannsen, 1932) on account of the frequently displayed flow banding (prefix rhyo-, from Greek *rhyox*, stream of lava). *Liparite* (Roth, 1861 see Johannsen, 1932) is synonymous with rhyolite but the term is now little used.

Varieties

Rhyolites have high SiO_2 (usually 68–77%), low normative color indexes (normative ol + opx + cpx + mt + il + hm ≤ 20) and normative plagioclase compositions $< \text{An}_{20}$ (Irvine and Baragar, 1971). They represent the evolved compositions of the subalkaline basalt-andesite-dacite-rhyolite magma series of either tholeiitic or calc-alkaline affinity. Peralkaline magmas can differentiate to either sodic or potassic alkali rhyolites such as *comendite* or *pantellerite*. Comendite can be chemically distinguished from pantellerite because it is less peralkaline and less Fe-rich. They both differ from rhyolite in having lower Al_2O_3 but higher TiO_2 and Na_2O .

The following are other varieties of rhyolite and related rocks.

Calcirhyolite—contains labradorite or bytownite as its accessory plagioclase.

Cantalite—a glassy variety.

Craignurite—a glass-rich variety of rhyodacite (Craignure, Mull, Scotland).

Dellenite—a rhyodacite.

Hakutoite—a quartz-poor alkali rhyolite, similar to taurite (Hakuto-San, Manchuria, China).

Ignispumite—characterized by lenticles and banding and transitional with ignimbrite; formed as an acidic foamy lava.

Khagiarite—a black hyalopantellerite characterized by a glassy microlitic groundmass exhibiting flow texture.

Krablite (*baulite*, *kraflite*)—originally identified as the mineral feldspar but is a rhyolite containing plagioclase enclosed in orthoclase phenocrysts together with smaller proportions of augite and quartz; occurs as ejecta (Mt. Krafla, Iceland).

Lithöidite—a cryptocrystalline rhyolite without phenocrysts.

Marloesite—phenocrysts of feldspar and pseudomorphs of mica after olivine in a groundmass characterized by glomerophytic texture; resembles pantellerite (Marloes, Wales).

Nevadite—contains abundant phenocrysts and only a small proportion of groundmass (Nevada, U.S.A.).

Obsidian—see **Volcanic glass**.

Perlite—see **Volcanic glass**.

Pyromeride—devitrified rhyolite with spherulitic texture.

Quartz latite—a rhyodacite.

Rhyobasalt—the effusive equivalent of granodolerite.

Rhyodacite—an effusive porphyritic rock intermediate in composition between rhyolite and dacite; quartz, plagioclase, and biotite (or hornblende) are the main phenocrysts; the groundmass may be fine-grained or glassy.

Saucyite—glassy and composed of sanidine phenocrysts in a groundmass of orthoclase microlites and minute crystals of biotite, augite, sphene, zircon and magnetite.

Taurite—a sodic rhyolite containing acmite; differs from *comendite* in having a spherulitic or granophytic groundmass.

Tordrillite—light colored, no mafic minerals and the same chemical composition as alaskite (Tordrillo Mtns. Alaska, U.S.A.).

Mineralogy

Mineralogically, rhyolites may contain phenocrysts of alkali feldspar, plagioclase feldspar, and quartz in a groundmass of alkali feldspar and quartz or a glass of equivalent composition. The alkali feldspar is typically anorthoclase ($\text{Ab}_{50}\text{Or}_{50}$) but may be sanidine in the alkali rhyolites. The plagioclase feldspar is typically oligoclase or albite. The subordinate mafic constituents may include augite, hypersthene, fayalitic olivine, biotite, magnetite, ilmenite,

and less commonly hornblende. Alkali rhyolites may contain such ferromagnesian minerals as aegerine-augite, riebeckite, and cossyrite.

Structural and textural features

Rhyolite magma has an average density of 2.4 g/cm³ at room temperature (basalt, 2.9 g/cm³) and a high viscosity at liquidus temperatures, which range from 735 to 950°C at atmospheric pressure (Carmichael et al., 1974). Rhyolitic magmas are considerably more viscous than basaltic magmas and hence rarely reach the surface. When they do, they tend to form short, thick lava flows or extrusive domes of rhyolite or obsidian. The rocks frequently exhibit contorted flow banding owing to the high viscosity. The more glassy varieties can exhibit perlitic fractures and spherulitic texture is sometimes produced by the supercooling of eutectoid melts. The extrusive domes (tholoids) can be up to hundreds of meters in diameter and they are frequently surrounded by a mantle of talus generated by autobrecciation. In subaqueous or subglacial settings, rhyolite eruptions generate fragmental products (hyaloclastite) often with the development of pseudopillows such as those from the Aso caldera, Japan, or Landmannalaugar, Iceland.

Ninety-five percent of rhyolitic eruptions occur as ash-flows. These highly explosive eruptions produce voluminous incandescent clouds of gas and rock particles that travel at great speeds, often surmounting considerable obstacles and covering areas of hundreds of km² to thicknesses of tens or even hundreds of meters. Welding of such hot particulate matter produces welded tuffs or ignimbrites. Pumice is a typical nonwelded product of such eruptions.

Genesis

Eruptions of rhyolite can occur in a variety of geotectonic settings, but they are particularly abundant at destructive plate margins such as North Island, New Zealand (Cole, 1979), the Central Andes and western Mexico (Cameron et al., 1980). In such settings they are typically the products of caldera-forming processes associated with mature stratovolcanoes.

Rhyolitic magmas are generated either by protracted differentiation of a basaltic parental magma or partial melting of pre-existing crustal rocks. Crustal melting is particularly important in destructive plate margin settings where the abundance of rhyolite is closely related to the presence or absence of continental crust. The remelting of plagiogranites in the ocean crustal layer, following rift jumping, may account for the abundance of rhyolites in Iceland rather than differentiation of basaltic magma. The

frequently reported bimodal association of basalt and rhyolite in the absence of intermediate compositions is often cited as evidence that crustal melting is important in the petrogenesis of rhyolite.

G. P. DURANT

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Cross-references: *Calderas and volcano-tectonic depressions; Endogenous dome; Explosive volcanic eruptions—classification; Igneous rocks—classification and nomenclature; Magma; Rift valley volcanism; Volcanic glass; Volcanic rocks; Volcanoes and volcanism.*

RIFT VALLEY VOLCANISM

The term *rift valley* was introduced late last century by J. W. Gregory to describe a large trough-like feature in E Africa formed by subsidence of the floor along nearly-parallel marginal faults. The subsequent recognition of morphologically similar features in other continental areas and in parts of the ocean floors led to the concept of a world rift system. The mid-ocean ridges and some of the continental rifts are sites of intense volcanic activity, but there are significant differences not only in the types of eruptions and the compositions of the volcanic rocks produced in each environment, but also in the structures present in the rifts themselves. Indeed, the troughs along the oceanic ridges are not strictly rifts by definitions that specify subsidence of pre-existing rocks as in the case of continental rift valleys (King, 1978; Mohr, 1982) but the term *oceanic rift* is now deeply entrenched in the literature.

Volcanism in oceanic “rifts”

The mid-ocean ridges are regions where new crust is generated during sea-floor spreading and

continental drift, and the oceanic basalts are thus by far the most voluminous products of rift volcanism. It will be appreciated, however, that information about these rocks can only be pieced together from samples recovered by operations such as dredging and drilling in the oceans, supplemented by studies of sporadic volcanic islands on or close to the ridge axes. The lavas along the oceanic rift zones are mainly tholeiites (silica-oversaturated basalts).

Modern geochemical investigations of the volcanic products have produced an impressive body of data, enabling fine distinctions to be made in parts of the ocean floors (Barberi et al., 1982). Consideration of characteristics such as trace-element patterns and isotopic ratios shows that the tholeiites erupted along active sectors of ridges are of a rather uniform type referred to as mid-ocean ridge basalts (MORB). In contrast, the lavas of transverse fracture zones and those of oceanic islands are more variable in composition and sometimes include alkali basalts (types with lower silica and higher contents of alkalis than tholeiites).

Many of the islands display rock series ranging from basic (basalt) to acidic (rhyolite), commonly attributed to processes or crystal fractionation at high levels within the crust. Different trends are determined by the compositions of the basalts, which may be chiefly tholeiitic (e.g., Iceland) or mildly alkalic (e.g., Easter, Ascension, and Bouvet islands). Subordinate strongly undersaturated alkaline rocks such as phonolites and nephelinites, similar to those that characterize some continental rifts, occur on a few islands (e.g., Gran Canaria, Kerguelen, and Hawaii), but these are all far removed from the ridge crests (see **Mid-ocean ridge and ocean-floor petrology**).

Volcanism in continental rifts

A widespread but mistaken view that all continental rift valleys are characterized by voluminous developments of alkaline volcanic rocks stems from a tendency to consider only the type example (the Gregory or Kenya Rift). In fact, there are great contrasts both in the amount of volcanic material in the various rifts, and in the compositions of the erupted rocks. Further misconceptions have arisen concerning relationships between tectonic and volcanic events. It is often incorrectly assumed, for instance, that major crustal doming always preceded volcanism and faulting, whereas there is mounting evidence to show that the first important tectonic event was downwarping of the crust to form a broad proto-rift depression (Mohr, 1982). Although it is not always appreciated, there is a fundamental difference

between pre-rift regional doming and tilts associated with late uplifts of rift shoulders (Williams, 1982).

A comparison of salient aspects of volcanism in continental rift zones can best be approached by first taking one example to illustrate features such as the compositional range of the volcanic rocks and their distribution in space and time.

Eastern rift system of Africa. Several well-preserved rift zones constitute the Afro-Arabian system, which extends some 6,500 km from Turkey in the north to Mozambique in the south. The African section has an eastern branch running from the Red Sea and Gulf of Aden southwards through Ethiopia and Kenya to Tanzania, and a western rift that passes through Uganda and joins the eastern system in Tanzania. Attention is focused here on the eastern branch because of the profuse volcanism along its entire length (Fig. 1; Baker et al., 1972).

In detail, the rifts display great structural complexity. Where they traverse the highland regions (the so-called "domes") of Kenya and Ethiopia, the rifts are clearly-defined topographic features 60–70 km across. The floors

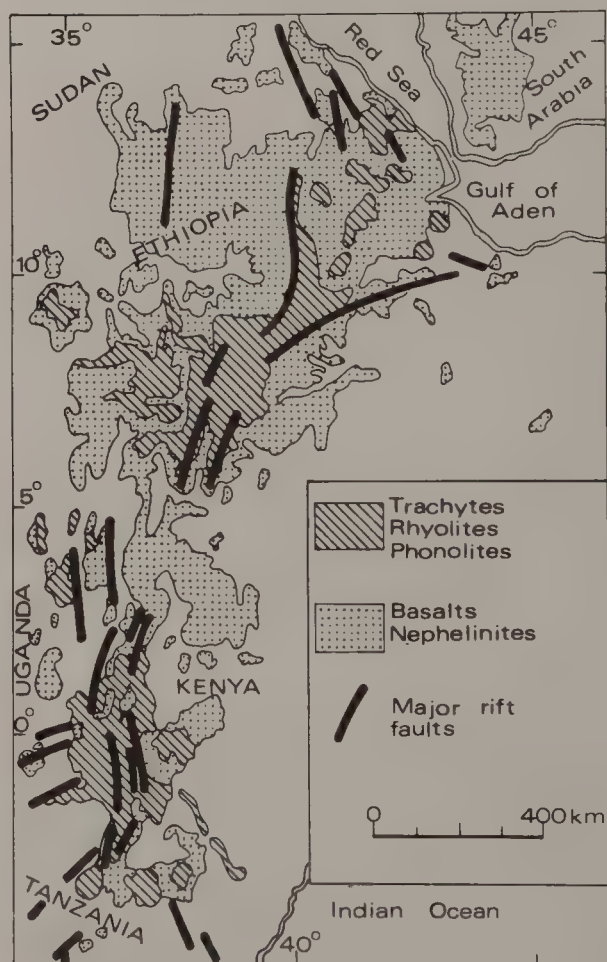


FIGURE 1. Volcanic rocks associated with the Eastern Rift System of Africa.

rise in these areas to maximum altitudes of 2,000 m in Kenya and 1,700 m in Ethiopia, commonly standing some 1,000–2,000 m below the level of the flanking ground. To north and south of these topographic culminations, the rifts splay out into broad depressions with less obvious boundaries.

The rift margins are often defined by step-faulted flexures, rather than by single major faults; only locally is there any approach to a simple graben structure. A major fault-scarp forming one side of the rift frequently stands opposite a margin dominated by downwarping of strata: in such cases, *half-graben* is an appropriate term to emphasize the structural asymmetry. The floors of the rifts are broken by numerous closely spaced faults, and there is a tendency for the most recent tectonic and volcanic activity to become concentrated in narrow inner zones. In some areas, the inner troughs are separated from the rift margin by complex tilted blocks or arch-like structures which expose sequences of older rocks (King, 1978).

Total vertical displacements exceeding 3 km occurred on some of the major faults, but the movements were spread over long periods so that accumulation of volcanic and sedimentary rocks more or less kept pace with subsidence of the rift floors. The rift depressions were therefore relatively subdued features throughout much of their long history spanning some 50 Ma. The present spectacular morphology was largely determined by a widespread episode of major faulting 1–2 Ma ago. The movements on the faults were mainly downward, so that deep rift infill was depressed below sea level, but late uplifts of the rift shoulders were important structural events.

The greatest thicknesses (at least 4 km) and the most complete successions of volcanic products and interbedded sediments lie buried within the rift zones, but volcanic rocks are not confined to the rift valleys (Fig. 1). At times, the rate of accumulation of infill outpaced subsidence of the depressions, and very fluid voluminous lava flows overtopped the developing rifts to spread considerable distances across the flanking ground. These flows coalesce with the products of widely distributed eruptions far beyond the confines of the troughs. The most impressive of these volcanic landforms are the massive central complexes such as Kilimanjaro and Mt. Kenya.

The volcanic rocks associated with the eastern rift system have a total volume of at least 500,000 km³ (Baker et al., 1972). Two broad compositional groups are distinguished in Fig. 1 to show the distribution of basic types (basalts, nephelinites) and salic varieties (trachytes,

rhyolites, phonolites). Basalts were erupted repeatedly both within and outside the rifts, and they are by far the most abundant rock types, with a volume exceeding 70% of the total. Nephelinites are volumetrically the least important volcanic rocks: they are largely restricted to central complexes near the Kenya–Uganda and Kenya–Tanzania borders. Trachytes and rhyolites (ca. 15% of total volume) occur mostly in or close to the rifts in both Kenya and Ethiopia, but flood phonolites (5–10%) are confined to Kenya. There are important differences between, for example, the ratios of basic to salic volcanic rocks in the Ethiopian and Kenyan provinces (6:1 and 1.5:1, respectively).

In Kenya and S Ethiopia, the volcanic products generally rest directly on Precambrian metamorphic rocks, but farther north they are widely underlain by Mesozoic sediments. Volcanism commenced in the north in the Eocene (lower Tertiary), and spread to the entire region of the present rifts by mid-Miocene times. Later Tertiary and Quaternary activity was nearly continuous along the full length of the rift system, persisting to the present day in the extreme north and south.

The oldest volcanic rocks are fissure basalts exposed on the Ethiopian plateaus. The ages are not precisely known, but the eruptions probably started about 45 Ma ago. Predominantly basaltic activity, some related to large shield volcanoes, was widespread in Ethiopia during the Oligocene and part of the Miocene, spanning the period 30–15 Ma ago. Basalts belonging to the same episode can be traced through a broad depression in northwestern Kenya, and then southwards along the rift margins almost to the equator (Fig. 2). Explosive nephelinite–phonolite–carbonatite volcanoes of the Kenya–Uganda border area were active over the period 25–17 Ma ago.

In later Miocene times (16–8 Ma ago) enormous volumes of phonolite were erupted in a broad downwarped depression in central Kenya, overflowing the margins to spread up to 400 km across the flanking ground that locally had considerable relief (Williams, 1978). At the same time, the focus of nephelinite–carbonatite activity shifted from the Kenya–Uganda border into the rift zone in S Kenya and N Tanzania where the oldest exposed volcanic rocks are nephelinitic lavas and pyroclastic deposits (15–12 Ma). Subsequent activity in this region was characterized by repeated episodes of similar strongly alkaline volcanism, and it continues to the present day in Tanzania. The Kenyan flood phonolites have no counterparts in Ethiopia, where their salic time-equivalents are mainly rhyolitic lavas and ignimbrites

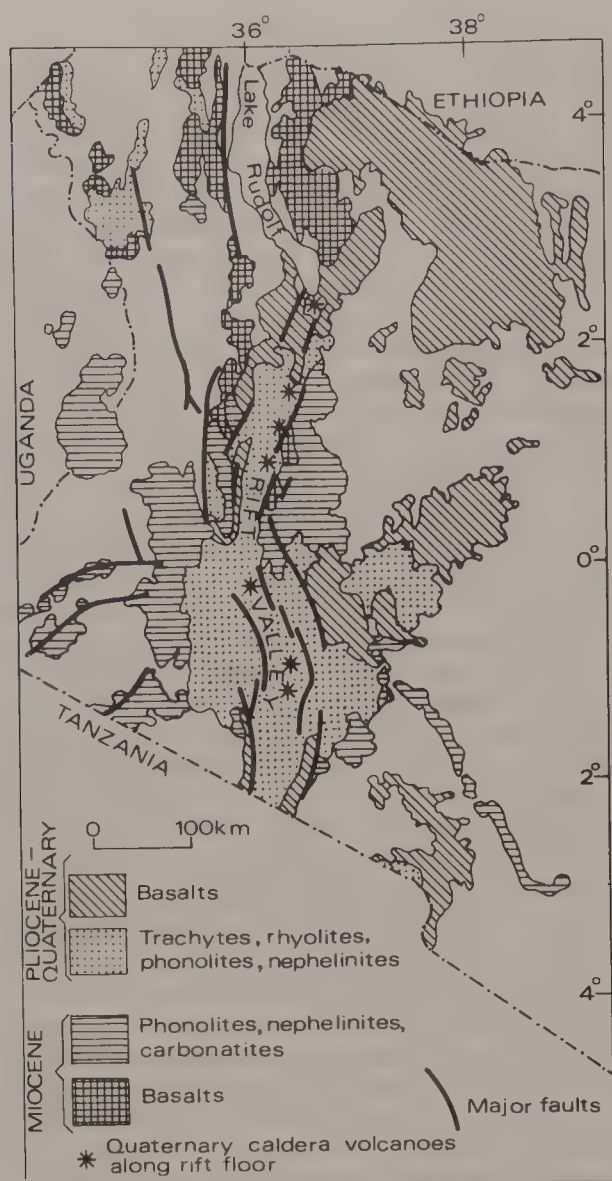


FIGURE 2. Distribution of rift-related volcanic rocks in Kenya.

(14–10 Ma) now exposed along the rift margins.

In Kenya, important changes in the compositions of the volcanic products followed a major episode of rift faulting in the late Miocene (7 Ma ago). Basaltic activity continued, but phonolites were largely superseded by trachytes that commonly form recognizable shield volcanoes (6–2 Ma) within the rift. These eruptions coincided with a second peak in the intensity of silicic volcanism in Ethiopia. In both regions, trachytic–rhyolitic ignimbrites and lavas locally infilled the rifts and spread across the flanks. Coeval Pliocene activity 100–150 km E of the Kenya rift produced the massive Mt. Kenya phonolitic volcano and some fissure basalts to the north.

Following major faulting in Pleistocene times, intense volcanism in the floors of the rifts

culminated in the development of chains of rhyolitic and trachytic caldera volcanoes, many of which remained active until very recent times. Basaltic fields fed by fissures are interspersed with the silicic volcanoes, becoming particularly prominent in the Afar depression in the extreme north, where some centers are still active. Great linear multicenter ranges 150–250 km east of the Kenya rift are composed almost entirely of Quaternary basalts. Mt. Kilimanjaro has a similar foundation, but the compositions of some of the lavas erupted during its unique later history are a reflection of proximity to young strongly alkaline centers in northern Tanzania.

The events described above show that in Kenya there was a marked eastward shift of the main areas of volcanism with time. Oligocene–Miocene activity was within the rift zone and to the west; Pliocene–Quaternary volcanism occurred in the rift floor and to the east. The pattern is rather different in Ethiopia, where there was less intense volcanism east of the rift depressions.

The Kenyan province shows the greatest diversity of composition (Fig. 3) but Na_2O consistently exceeds K_2O . The rocks have distinctive mineralogical compositions. Basalts contain plagioclase feldspar, pyroxene, and commonly olivine; some bear traces of nepheline, analcime, or zeolites and are better classified as basanites. Trachytes are characterized by alkali feldspar together with sodic amphiboles and pyroxenes; some contain olivine. Quartz-bearing varieties are common, and with increasing silica content these grade into sodic rhyolites. As the name implies, trachybasalts are mineralogically and chemically intermediate between trachytes and basalts; terms such as mugearite and hawaiite are often used to describe these rocks. Phonolites (nepheline-bearing trachytes) are divided into two fields in Fig. 3. Lavas of type A contain alkali feldspar, nepheline, analcime, and sodic pyroxenes and amphiboles; biotite is not uncommon and some of the rocks are olivine-bearing. Trachyphonolites are very similar, but nepheline is less conspicuous. Most phonolites of type B are characterized by the presence of sphene and lack of amphiboles. They are commonly associated with nephelinites that are essentially nepheline–pyroxene rocks containing combinations of minerals such as olivine, biotite, perovskite, melilite, analcime, and zeolites. Silicate minerals make up the bulk of rocks defined above, but carbonatites are composed almost entirely of carbonates: they do not appear in Fig. 3 because they have very low silica contents.

The basic volcanic rock types (basalts,

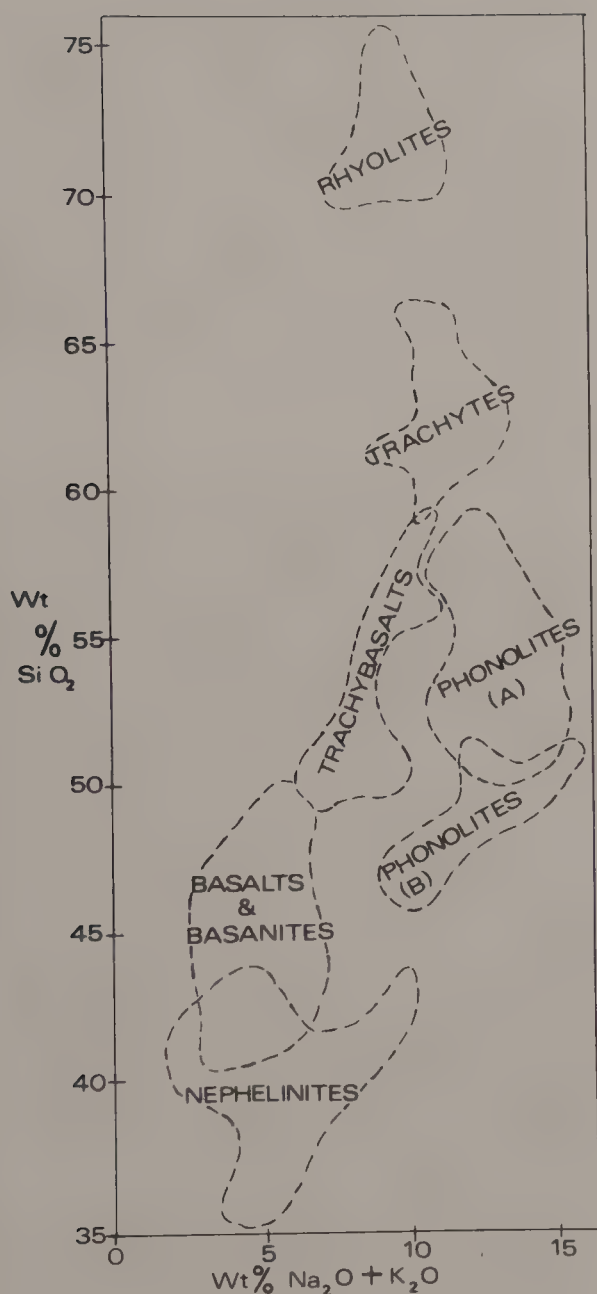


FIGURE 3. Alkalies vs. silica diagram for Kenyan lavas, based on more than 200 samples; $\text{Na}_2\text{O} > \text{K}_2\text{O}$ in all rocks.

basanites, nephelinites) span widely ranging values of silica and alkalies. More elaborate diagrams than Fig. 3 bring out temporal and spatial trends. In the northern part of the Kenya rift, for example, basic lavas show a trend of decreasing alkalinity and silica-undersaturation with time. The older lavas are mainly alkali olivine basalts, whereas the Quaternary flows are "transitional" basalts showing affinities with tholeiites.

Evidence similar to that presented in Fig. 3 was used in early accounts to distinguish two genetic series—one strongly alkaline, embracing nephelinites and phonolites, and the other more mildly alkaline, including basalts, trachybasalts,

trachytes, and rhyolites. Later it became apparent that other trends are also represented, and that some of the volume relationships require explanation. The most challenging petrogenetic problems are those concerning the origins of phonolites, trachytes, and rhyolites (Baker et al., 1972; Mohr, 1982).

Comparison with other continental rifts. Attempts to match the compositions and volumes of the Kenyan volcanic rocks with those reported from similar structural environments elsewhere are generally unrewarding (Williams, 1982). Profuse volcanism in the active Rio Grande rift of the southwestern United States invites comparison with Kenya, but the calc-alkaline suites of the former and the nephelinite-phonolite associations of the latter leave the two regions with little in common other than major occurrences of alkali basalts. Although much older (Permian), the Oslo graben of Norway more closely resembles the Kenya rift in having abundant alkali basalts, some strongly undersaturated basic lavas, and substantial proportions of trachytes and rhyolites. Moreover, the voluminous rhomb porphyries (lavas with large rhomb-shaped feldspars) of the Oslo region bear some resemblance to the Kenyan flood phonolites in features such as uniformity of texture and fluid behavior during emplacement, but they are less alkaline and frequently contain quartz.

Volcanism played only a minor role in the evolution of some rifts. Although it follows that processes such as evacuation of large magma chambers cannot be invoked in seeking a fundamental mechanism for graben formation, an absence of volcanic infill is undoubtedly responsible for the survival of some very deep troughs now occupied mainly by water (e.g., Lake Baikal, Siberia, U.S.S.R.; Lake Tanganyika). Moreover, paucity of volcanic rocks in some rifts and their profusion in others must reflect differing thermal conditions in the underlying asthenosphere (Mohr, 1982). The scale of volcanic activity thus provides a meaningful basis for a classification of continental rifts. Two broad groups can be distinguished (Barberi et al., 1982). The Kenya-Ethiopia-Afar rift system takes pride of place in a high-volcanicity group: indeed, it is probably the only true representative. The European Rhine graben, the Baikal rift of Siberia, the western rifts in E Africa, and the Dead Sea rift in the Middle East all belong to a low-volcanicity group. The Rio Grande rift of the United States should perhaps be placed in a separate category indicating moderate volcanicity. With a ratio of volume of volcanic products to length of rift of about 15:1, the Rio Grande is certainly no match for the eastern

branch of the African system, where the ratio is nearly 300:1. The subdivision is more difficult to apply to palaeorifts because deep levels of erosion usually result in uncertainty about the distribution and thickness of volcanic rocks; intrusive complexes provide some guidance.

Low-volcanicity rifts are characterized by rather wide variations in the chemistry of basic lavas, and by somewhat restricted developments of salic products that generally display only limited compositional ranges. Strongly alkaline undersaturated basic rocks (nephelinites, basanites) usually occur in the widely scattered volcanic fields in these rifts.

The spectacular differences between low- and high-volcanicity rifts are attributed to greater rates of extension and more marked attenuation of lithosphere along the latter (Barberi et al., 1982), so that the main factor determining the scale of volcanic activity is seen to be the extent to which asthenosphere penetrates the lithosphere. The predominantly strongly alkaline character of rocks in the low-volcanicity rifts reflects the great depths of mantle melting beneath these zones. In rifts displaying more voluminous volcanic products, diapiric uprise of mantle leading to progressively shallower levels of melting is an attractive hypothesis to explain local trends such as decreasing alkalinity of lavas with time. However, elaborate models are required to account for the full pattern and history of volcanism.

L. A. J. WILLIAMS

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Cross-references: *Alkaline rocks—undersaturated; Basalt; Carbonatite; Feldspathoidal rocks—genesis; Igneous rocks—tectonic setting; Leucitite; Mid-ocean ridge and ocean-floor petrology; Nephelinite; Peralkaline igneous rocks; Phonolite; Rhyolite; Trachyte; Volcanic rocks.*

RING DYKE

Major intrusions that are arc-shaped or annular in plan, with either vertical or steeply inclined sides, are termed *ring dykes (dikes)*. They occur singly or in concentric or overlapping sets. Individual ring dykes often vary only slightly in width along their course, but in other cases there are wide variations. Great differences occur in the width of different ring dykes, even in the same intrusive complex, some being as narrow as 100 m while others exceed 1.5 km. The broader examples sometimes have internal gradational variations indicating intrusion by a succession of magmatic pulses of slightly differing composition or physical condition. The diameters of ring dykes or ring intrusion complexes are highly variable but do not usually exceed about 15 km even where there has been migration of the center of intrusion. Emplacement may be brought about by one or more of several mechanisms,

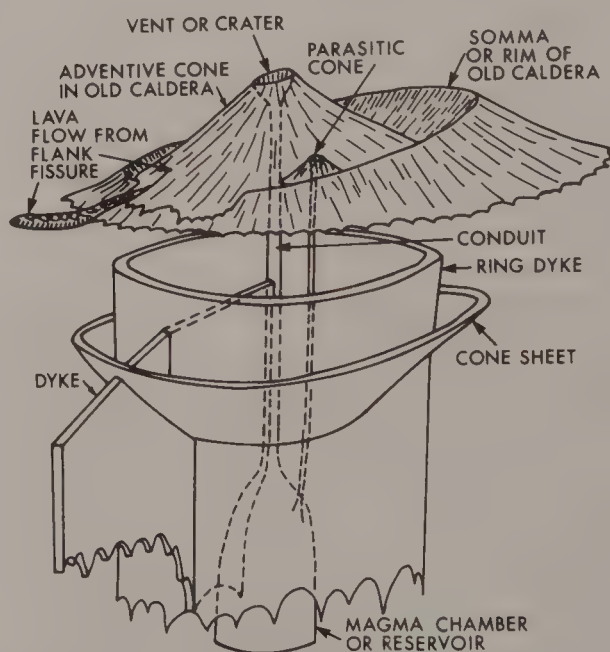


FIGURE 1. Schematic representation of relationships of a ring dyke and a cone sheet to surface expressions of volcanism.

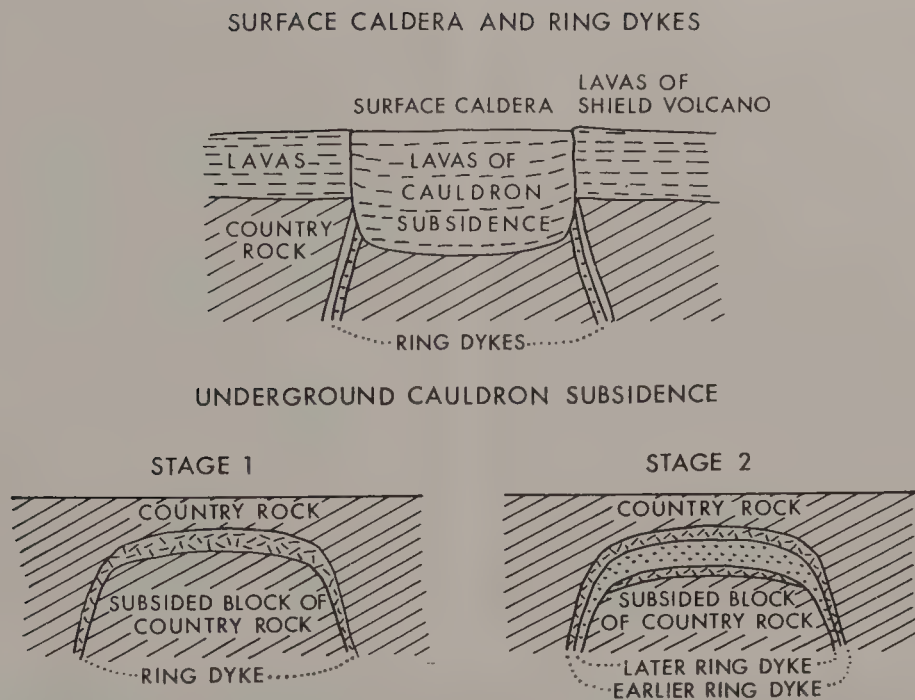


FIGURE 2. Diagrammatic cross-sections of cauldron subsidences. (After Richey, 1961.)

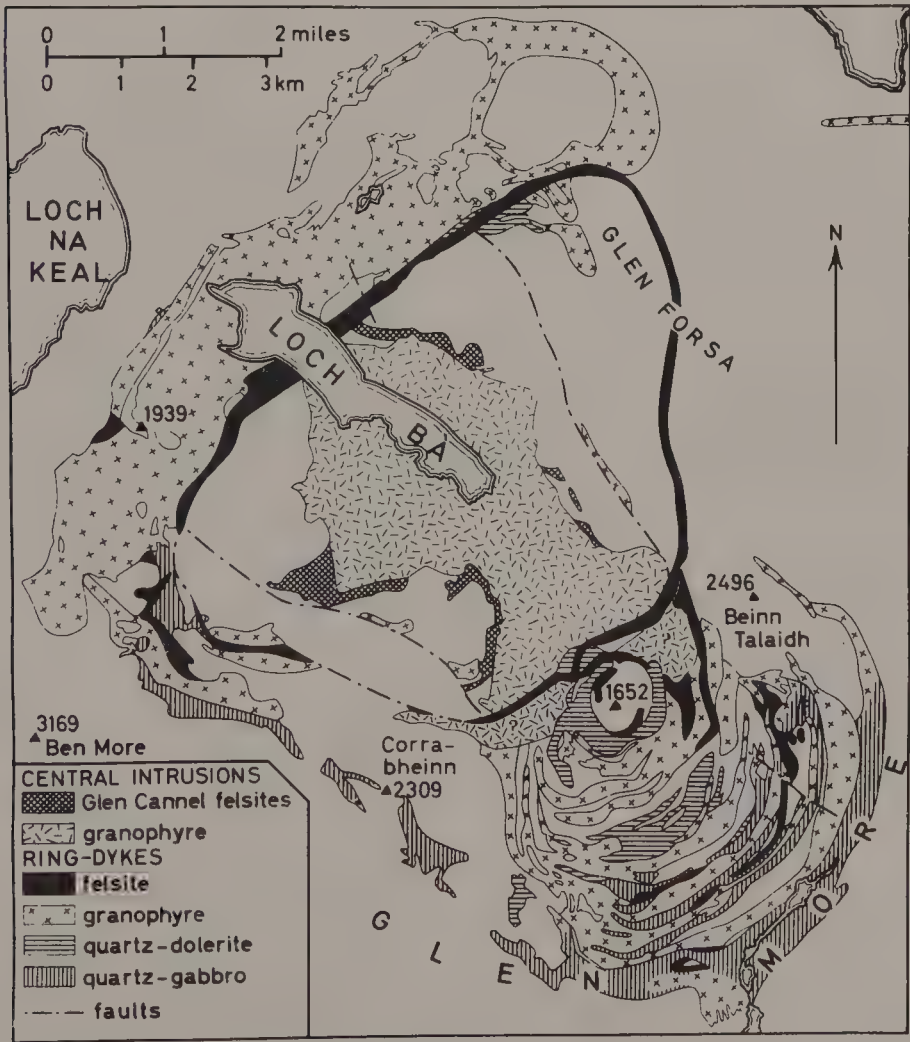


FIGURE 3. Loch Bà ring dyke and other ring intrusions in Mull, Scotland; the host rock is basalt. (After Bailey et al., 1924.)

including cauldron subsidence, stoping, and gas coring.

In cauldron subsidence the space for intrusion may be provided by the subsidence of a central block within a ring-shaped or arcuate fault with a steeply outward-dipping inclination. The block subsides into a magma chamber and the magma is injected up the ring fracture, where it solidifies to form a ring dyke (Figs. 1, 2), e.g., the Loch Bà ring dyke, Mull, in the British Tertiary Volcanic Province, which has been intruded along a steeply outward-dipping fracture within which there has been subsidence of over 900 m (Fig. 3). The resulting dilation from this subsidence is sufficient to provide enough space for the associated felsite intrusion. A similar example occurs in the Oslo region of Norway (Fig. 4) where there was subsidence of ca. 700 m within the ring fracture. In both these

instances the ring dyke is narrow in proportion to its diameter. In other instances much thicker intrusions, and intrusions of variable thickness, would require an impossible degree of subsidence of the central block to achieve sufficient dilation. Also, in some such cases it can be demonstrated that the ring fracture is vertical or even steeply inward-dipping, hence the simple cauldron subsidence mechanism cannot account for emplacement, and indeed could even inhibit emplacement of magma in a ring fracture.

An example in which the contacts are vertical occurs in the Northern Ring Complex of the Nuanetsi Igneous Province, Zimbabwe (Stillman, 1970) where a single ring dyke of porphyry partly encloses a layered gabbro mass intruded into Karroo basalts. In this case subsidence of the central block varies from zero in the east to

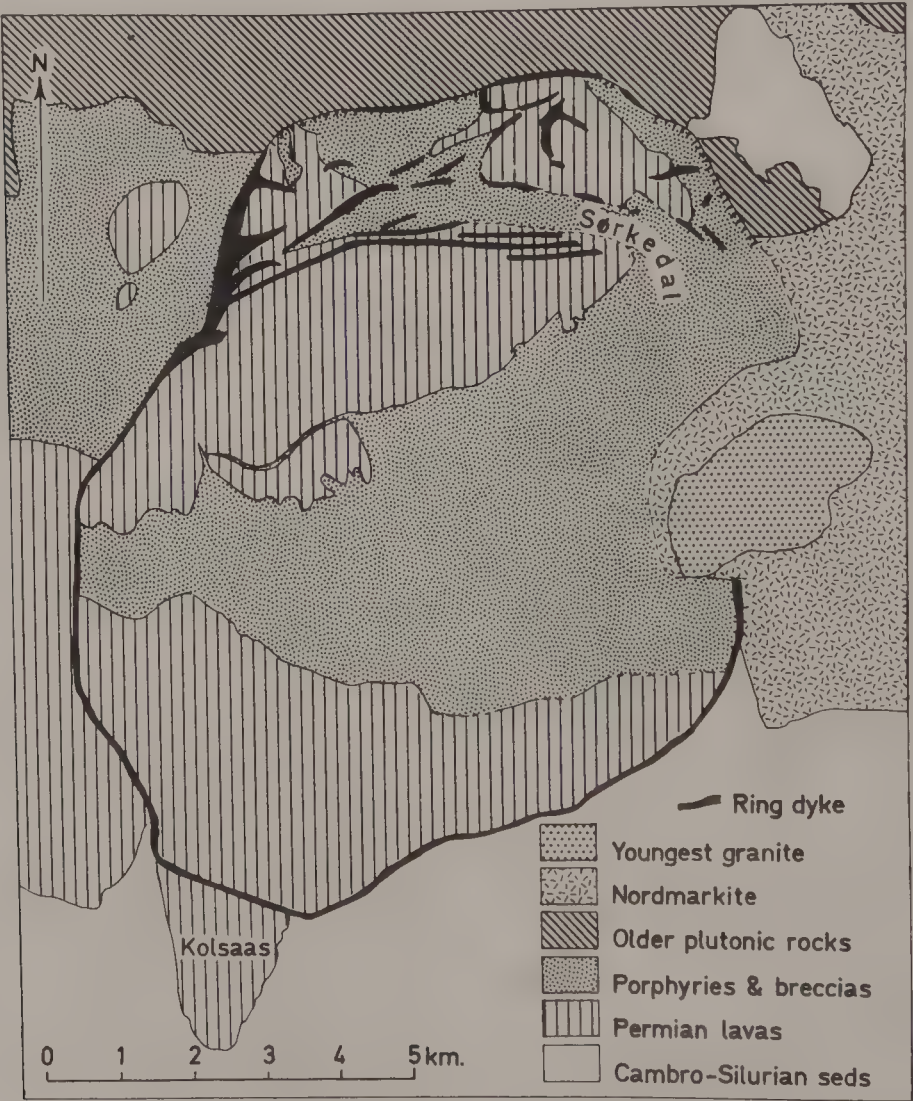


FIGURE 4. The Baerum Caldera, Oslo region, Norway, of Permian age emplaced in Lower Palaeozoic sediments. The ring fracture has been intruded by a ring dyke of syenite porphyry that has been partially obliterated by later intrusions of nordmarkite and granite. (After O. Høltedahl and J. A. Dons (1966), 1:50,000 geological map of Oslo and district, Oslo: Universitetsforlaget.)

ca. 1200 m in the southwest. This uneven tilting has been aided by the fragmentation of the block along many fault planes. It is considered that the subsidence took place at the time of eruption of large quantities of silicic tephra that formed extensive ignimbrite flows, the rapid withdrawal of magma from below leading to formation of a caldera at the surface. The ring fracture produced during the formation of the caldera is vertical or near-vertical; thus, there can have been no appreciable dilation effect to account for the emplacement of the ring dyke. Moreover, the intrusion contains abundant xenoliths of Karoo basalts, indicating that the space for emplacement was made by displacing the country rock. A possible mechanism for this would be by coring by highly fluidized magma at the time of eruption of the ignimbrites.

The surface expression of this type of ring structure may be comparable to the Valles Caldera, New Mexico, U.S.A. (Smith, et al., 1961) although in that instance (Fig. 5) initial subsidence of the central block appears to have been followed by resurgence or elevation of the caldera floor. Evidence of such resurgence can be seen in some more deeply eroded ring complexes such as the Tertiary volcanic center on the island of Arran, Scotland (King, 1955). The close genetic relationship between intrusive ring complexes and volcanic calderas seems to be confirmed by the similar range of diameters of both phenomena (cf., Fig. 1).

Ring dykes also occur in concentric sets, adjacent intrusions either being in contact with each other or in some instances separated by narrow screens of older rock. Examples of the latter case occur in Mull, e.g. northwest of Loch

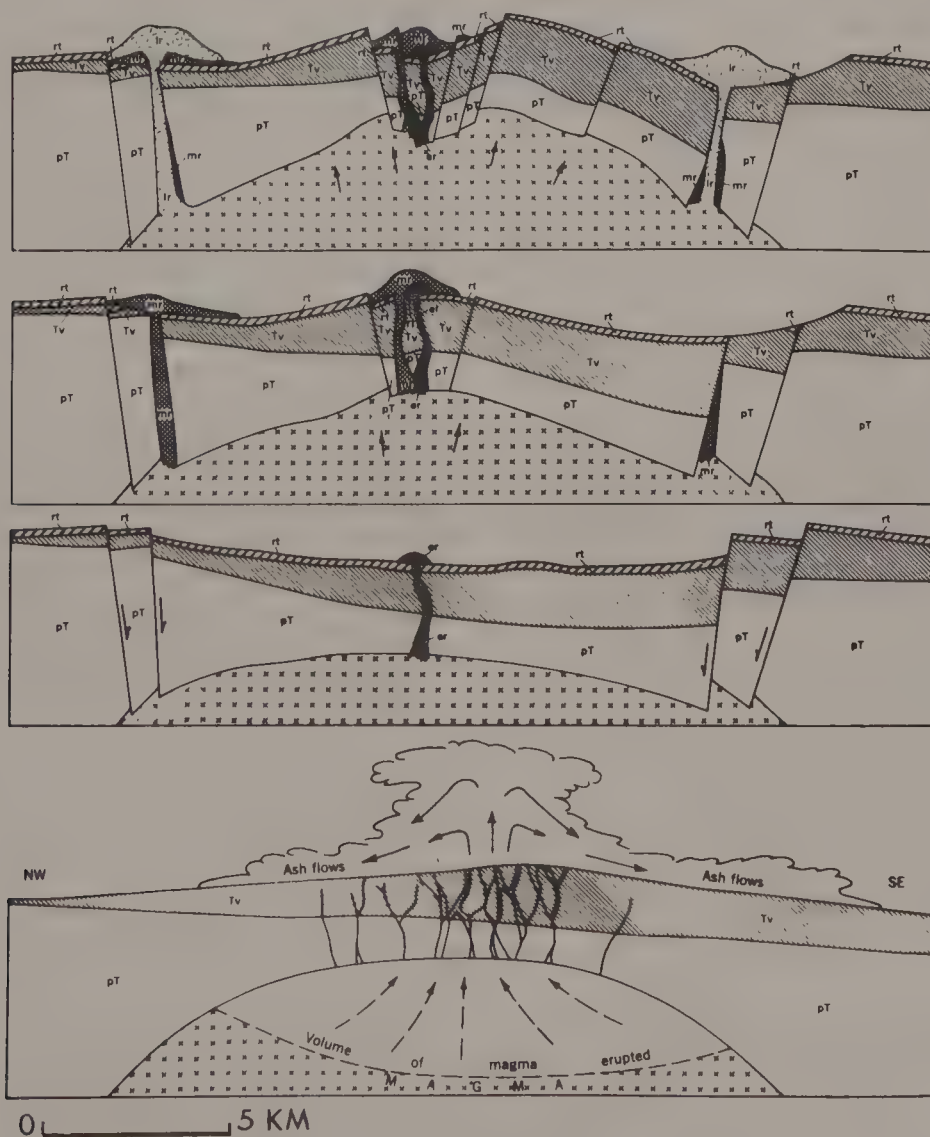


FIGURE 5. Diagrammatic cross-section of Valles Caldera, New Mexico, U.S.A. The ring intrusion took place after initial subsidence of the caldera, magma being injected into the ring fracture during later doming of the caldera floor. (From Smith et al., 1961.)

Bà, where a screen of Tertiary plateau basalt occurs between two granophyre ring dykes (Fig. 3). The scaling off of screens in this way could be another way in which extra space is made for the intrusion of thicker ring dykes at deeper levels as the result of a decrease in the diameter of the central block. Space may also be made as the result of stoping of large blocks of country rock that subside into the magma as it is intruded.

Other examples of multiple ring dyke complexes include the Ardnamurchan ring complex, Scotland (see **Cone sheets** for map), where three successive phases of multiple intrusion may have resulted in the emplacement of overlapping sets of ring dykes, mainly gabbroic in composition (Richey, 1961). The occurrence of basic rocks in this type of structure, however, is not common. Elsewhere in the British Tertiary Volcanic Province there are granitic ring dykes and in other areas throughout the world ring dykes are predominantly composed of silicic or alkaline rocks: examples include the Virginia Dale Precambrian complex, Colorado-Wyoming, U.S.A. (Eggler, 1968), with intrusions of granite and quartz monzonite, and the ring complexes of Nigeria, where individual intrusions, including granites and syenites, occur with diameters of 9–25 km (Jacobson et al., 1958).

Other structures often found associated with ring dykes include roof sheets, cone sheets, and linear and radiating dyke swarms (Fig. 1). Roof sheets are formed as the result of underground cauldron subsidence (Fig. 2b), the thickness depending on the amount of subsidence of the central block; they are probably formed also in some cases where a surface caldera is involved as a result of horizontal fracturing in the subsiding block. A variety of minor intrusions in the form of cone sheets and dykes are commonly but not invariably associated with ring complexes. Some of the best examples occur in the British Tertiary Volcanic Province where each of the major centers has its associated dyke swarm. The close connection between major central volcanic eruptive centers

and widespread fissure eruptions, on the one hand, and subadjacent major and minor intrusions on the other, can be seen by comparing the Tertiary and Recent volcanism in Iceland to the older and more deeply eroded centers in the Hebrides, Scotland (Walker, 1964, 1975).

J. G. MACDONALD

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Cross-references: *British Tertiary Volcanic Province*; *Calderas and volcano-tectonic depressions*; *Cone sheets*; *Dyke (dike)*; *Emplacement—modes*; *Hypabyssal rocks*; *Plutonic rocks*; *Tephra*; *Volcanoes and volcanism*.

ROCK-FORMING MINERALS—See
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S

SCHIST

A *schist* is a medium- or coarse-grained metamorphic rock characterized by a parallel or subparallel arrangement of the constituent platy, fibrous, and prismatic minerals such as micas, chlorite, talc, certain amphiboles and other minerals. Its most striking feature is the planar, often linear but ubiquitous *schistose structure* (*schistosity*) which imparts on the rock a fissility or a tendency for the rock to split along certain more or less parallel planes. A foliation (not synonymous with schistosity) or lamination of the minerals into parallel thin layers is generally well developed. The planes of schistosity are commonly referred to as *s-surfaces* (Sander, 1930).

Although the term schist encompasses many rocks (i.e., biotite schist, green-schist, lamproschist, para-schist), most are regional metamorphic derivatives of pelitic (aluminous) sediments such as clays, slates, mudstones, and to a lesser extent basic igneous rocks. Metamorphic crystallization and deformation played a major role in their development, with the latter causing many minerals to grow or rotate into preferred orientation directions.

Indication of the nature of schists is generally given by listing the prominent minerals present (e.g., quartz-garnet-biotite schist), but, in places, particular names have been given.

Ollenite—a type of hornblende schist with abundant epidote, sphene, and rutile and accessory garnet (Col d'Ollen, Piedmont, Italy).

Prasinite—a green schist with hornblende, chlorite, and epidote present in approximately equal proportions.

Sismondinite—sismondine (Mg-chloritoid) is the main constituent mineral.

Staurolite—a mica schist, commonly garnetiferous, characterized by the presence of porphyroblasts of staurolite.

During their formation, most schists pass through intermediate phases such as slate and phyllite. The changes may be envisaged as continuous grain growth and a progressive change in the mineralogy with increasing metamorphic grade. With further increase in grain size, the onset of partial melting or segregation during high-grade or ultrameta-

morphism, or an increase in the quartzofeldspathic constituents in the rock, a schist may pass into a gneiss. Considerable ambiguity remains, however, as to where to draw the boundary between phyllite and schist on the one hand and schist and gneiss on the other.

Large crystals (porphyroblasts) of Al-silicates such as garnet, kyanite, andalusite, and sillimanite often appear at some stage in the development of (aluminous) schistose rocks (Fig. 1). In recent years, petrographic work on schists has often focused on petrofabric studies,



FIGURE 1. Camera lucida drawing of a portion of a rock thin section shows the relationship between an earlier (S_1) and a later (S_2) schistosity in a polyphase deformed schist. S_1 is seen as a relict internal schistosity in the porphyroblasts of garnet and staurolite (Si) in the form of planar trails of fine quartz inclusions (Q) and as microfolded small mica flakes in the matrix. S_2 is defined by coarser mica flakes that have grown during a second metamorphic event along the hinges of microfolds of S_1 .

the growth in time of porphyroblasts, and their relationship to schist matrix and major structures. Together with the intricate microstructures present in many schists, such studies have permitted the elucidation of the varying conditions of temperature and pressure during prolonged metamorphism or orogenesis.

A. E. SHIMRON

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Cross-references: *Gneiss*; *Metamorphic mineral growth*; *Phyllite*; *Polyphase metamorphic mineral growth*; *Porphyroblast*; *Regional metamorphism*.

SERPENTINITE

Serpentinite is a soft, compact, pale green to greenish-black rock consisting wholly or largely of serpentine minerals—antigorite, chrysotile, or lizardite (allomorphs of $\text{Mg}_3\text{Si}_2\text{O}_5(\text{OH})_4$). These minerals are usually produced by the hydrous alteration of olivine, orthopyroxene, and, to a lesser extent, clinopyroxene. Most serpentinite bodies are altered ultramafic igneous rocks, and because they are the products of alteration they have a gradational relationship with peridotite. Less commonly serpentinite may be the product of the hydrous retrogressive alteration or metamorphism of forsterite or diopside marbles, and in the case of komatiites (ultramafic lavas) they result from the breakdown and alteration of high-Mg volcanic glass.

Serpentine minerals are an essential component of kimberlite (mica peridotite) and some kimberlites are so hydrated as to have a serpentinite matrix.

Serpentinities are rarely monomineralic, and usually contain two mineral parageneses: the relict assemblage consists of corroded remnants of the original ultramafic rock—olivine, orthopyroxene, clinopyroxene and spinel (usually chromite, picotite, or pleonaste)—and the secondary assemblage consisting of one or more of the allomorphs, a spinel (usually magnetite or ferrichromite), talc, and chlorite. Other phases often, but not always present, include anthophyllite, cummingtonite, tremolite-actinolite, carbonates (magnesite, dolomite, or calcite), sulfides (pyrite or pyrrhotite), and oxides (hematite, periclase, and brucite). Clay minerals may be important, these being largely vermicul-

ite. Where talc dominates the assemblage, the rock is termed *soapstone* or *steatite*. A type dominated by compact, asbestiform tremolite (nephrite) is sometimes called *jade*.

Petrographically serpentinites are diverse. In many the alteration process has been one of static recrystallization, and the original textures of the ultramafic rock are pseudomorphed by the serpentine minerals. Particular pseudomorphs are quite specific, with individual names—e.g., a metallic, fibrous amphibole replacement of orthopyroxene is termed bastite, and a pyribole mineral that often replaces clinopyroxene is termed uralite.

Bowenite (*tangawaite*, *tangiwaite*, *tangiwai*—Maori) is the name given in New Zealand to a particular type of serpentinite composed of a dense felt-like aggregate of serpentine fibers with flakes of talc, grains of chromite, and occasional patches of magnesite; occurrence is as veins in a foliated rock containing the same minerals but having talc as the major constituent.

Massive serpentinite can often be devoid of any relict features, consisting entirely of a medium- to fine-grained felted mass with semiregular cross-cutting veins, referred to as mesh and net texture. Such an assemblage is usually dominated by chrysotile and lizardite. Deformed serpentinite bodies are usually foliated to varying degrees, with antigorite being more important, and with talc and chlorite often defining the foliation. A complete spectrum exists between serpentinite with spaced foliae and serpentine-chlorite-talc schist.

The occurrence of serpentinite follows the pattern of ultramafic rocks, the largest occurrences being in ophiolite complexes and Archaean greenstone belts. Serpentinite also forms a variable part of layered igneous complexes and possibly comprises large parts of the deep ocean crust.

Economic interest centers around serpentinites as a source of industrial materials. Both the whole rock and chromite segregations are worked as a source of refractory raw material. Asbestiform varieties of anthophyllite, chrysotile, and tremolite are all widely exploited, particularly in the maritime states of Canada and New England, U.S.A. Talc is perhaps the most important raw material, particularly for the paper industry in Scandinavia. Podiform chromite segregations form a minor, though important source of metallurgical grade Cr being worked for this purpose in Cuba, the Philippines, and Albania; elsewhere these deposits are exploited for the refractories industry.

From earliest time, and today in many

primitive cultures, soapstone and serpentinite are valued as an ornamental stone. The Inuits of N America and the Lapps of N Europe traditionally manufacture artifacts in both materials, while the New Zealand Maoris have a long and distinguished tradition of using nephritic jade. Serpentinite and serpentine marbles are still highly valued as ornamental, monumental, and facing stone in buildings and monuments.

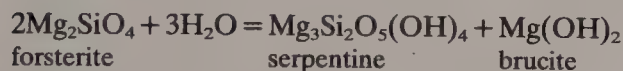
ADRIAN F. PARK

Cross-references: *Alpine mafic magma stem; Intrusion—layered; Kimberlite; Oceanic crust; Ophiolite; Peridotite; Steinmann Trinity; Ultramafic lavas; Ultramafic rocks.*

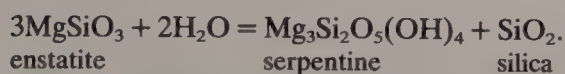
SERPENTINIZATION

Serpentinization is the process whereby anhydrous ferromagnesian silicate minerals are converted to hydrous ferromagnesian silicate minerals. This is shown where ultramafic rocks are converted to serpentinite.

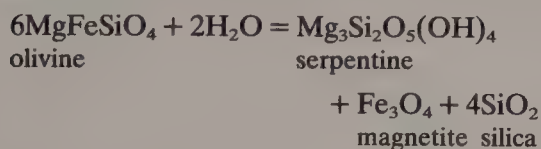
The following are the basic mineral reactions involved



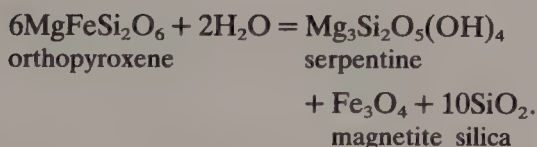
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Where the anhydrous phases are Fe-bearing, the following are the mineral reactions involved:



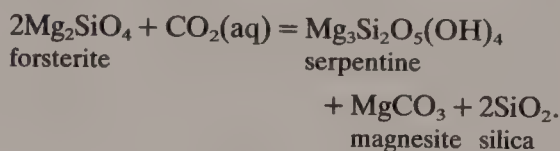
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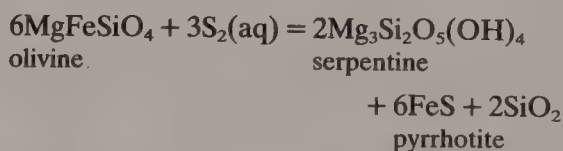
These transformations are matched by a radical change in the physical properties of the whole rock, e.g., density changes from ca. 3.2 gm/cm³ for an ultramafic rock to 2.6 gm/cm³ for serpentinite. This either represents a 30% mass loss or a 35–40% volume increase during serpentinization. For many years this formed the background to a constant-mass vs. constant-volume controversy (cf., Thayer, 1966; Hostetler et al., 1966), but it is unlikely that one process is universally

applicable, and in individual examples either case can be made, even in different parts of the same serpentinite body. In general, a case can be made for addition and removal of material during serpentinization. Volatiles, principally H₂O, are added, along with CO₂, S, Cl, and F, while Mg, Fe, and Si may be lost. In examples of bodies being actively serpentinized spring water carries MgOH⁺, Mg²⁺, and H₃SiO₄[−] as well as Fe salts in solution, often near saturation (Barnes and O'Neill, 1969). In the geological record, many serpentinites are surrounded by haloes of chemically altered (metasomatized) country rock to which Mg, Fe, and Si appear to have been added.

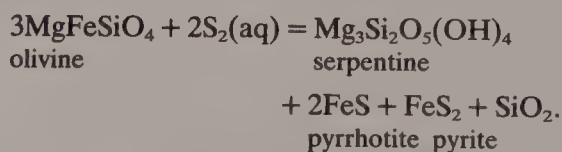
Which elements are added and removed depends to a large extent upon the chemistry of the aqueous solutions involved in serpentinization, as well as the *P–T* conditions at which the process takes place. The presence of CO₂ both leads to the generation of different phases (mainly carbonates) and interferes with the stability of silicates. Thus, magnesite may be an important phase:



Where sulfurous waters have a radical effect on the serpentinization of iron-bearing species the following reactions occur:



and



Cl and F are also important components of the aqueous medium but do not form independent phases. Instead, most analyses of serpentine minerals indicate the presence of F and Cl: the general formula for serpentine (Mg₃Si₂O₅(OH, Cl, F)₄) reflects this.

The hydration reactions that convert anhydrous ferromagnesian minerals to serpentine can proceed below 500–600°C, but most examples for which temperatures have been directly determined, or experimentally simulated, indicate a somewhat lower range (<300°C). Several examples of active, modern serpentinizing systems involving groundwater have been documented (Barnes and O'Neill,

1969; Campbell, 1975), while many other cases occur in which serpentinization formed a facet of a regional greenschist facies metamorphic regime. The assemblages chrysotile + magnetite + brucite and antigorite + magnetite, both with or without lizardite, appear to represent the low- and high-temperature ends of this spectrum. The presence or absence of brucite and magnesite reflects the fluid compositions rather than P - T conditions (Moody, 1976).

These serpentine-dominated assemblages are the products of retrogression, and where there are superimposed products of prograde greenschist metamorphism, phases like Mg-chlorite appear, probably indicating the breakdown of chrysotile and spinel. This process involves dehydration and places it into the realm of deserpentinization and prograde alteration.

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Cross-references: *Deserpentinization*; *Greenschist facies*; *Metasomatism*; *Peridotite*; *Serpentinite*; *Ultramafic rocks*.

SHOCK METAMORPHISM

Shock metamorphism is the process by which materials are permanently changed as a result of the passage of high-pressure shock waves. Detailed investigation of shock metamorphic effects in geologic materials dates from about 1960 (Chao et al., 1960), but the field of study includes such long-known phenomena as Neumann bands in iron meteorites and the glassy plagioclase phase, *maskelynite*, first described in the Shergotty achondritic meteorite by Tschermak (1872). Shock effects have been recognized in many meteorites (Dodd, 1981), in returned lunar samples (Engelhardt et al., 1970), and in terrestrial rocks both from large meteorite

craters (>90 m across) and from yet larger eroded impact structures (astroblemes). They are interpreted in all cases as prime evidence for hypervelocity impact (Chao, 1967), the only recognized natural means of generating high shock pressures (French and Short, 1968). Comparable effects produced by shaped explosive charges and high-velocity impact experiments in the laboratory and, on a larger scale, by buried nuclear explosions indicate that in rocks the most distinctive transformations occur at pressures in the range 100-600 kb, pressures much higher than those of explosive volcanism or tectonic deformation (Stöffler, 1972).

Shock waves and Hugoniot relationships

Shock waves are compression waves propagated at velocities exceeding the speed of sound in the medium specified. Characteristically, the compression is transient and confined at a given instant to a relatively narrow region that is strained at rates of 10^6 - 10^9 /sec. (Hörz, 1971). Such strain rates exceed by 10^{20} times or more those typical of deformation during regional metamorphism.

The changes in pressure (p), specific volume (V) and internal energy (E) across the shock-wave front are expressed by the Rankine-Hugoniot relation:

$$E_s - E_o = \frac{1}{2}(V_o - V_s)(p_s + p_o)$$

where (o) and (s) denote original and shocked states, respectively. This relation when combined with the equation of state of any material gives the Rankine-Hugoniot curve (or Hugoniot) which is the locus of all pressure-volume-energy states in the shocked material (Fig. 1).

For most geologic materials the Hugoniot has a cusp corresponding to the dynamic or Hugoniot elastic limit (HEL), below which dynamic compression does not produce permanent structural changes. Further discontinuities in the Hugoniot curve above the HEL mark changes to high-density polymorphs. The shock process results in an increase in internal energy that is partly retained as residual heat on decompression.

Products of shock metamorphism

Many products of shock-wave action, such as fracturing, brecciation, pseudotachylite-veining, and melting of rocks are megascopically similar to those of volcanic and tectonic activity. Unique to shock-metamorphosed rocks are shatter cones, which show striated, conical fracture surfaces with parasitic, horsetail-like partial cones on the master cone (Fig. 2a). The master cones are 0.05-1 m high with apical angle ca. $90^\circ \pm 30^\circ$, and are commonly incom-

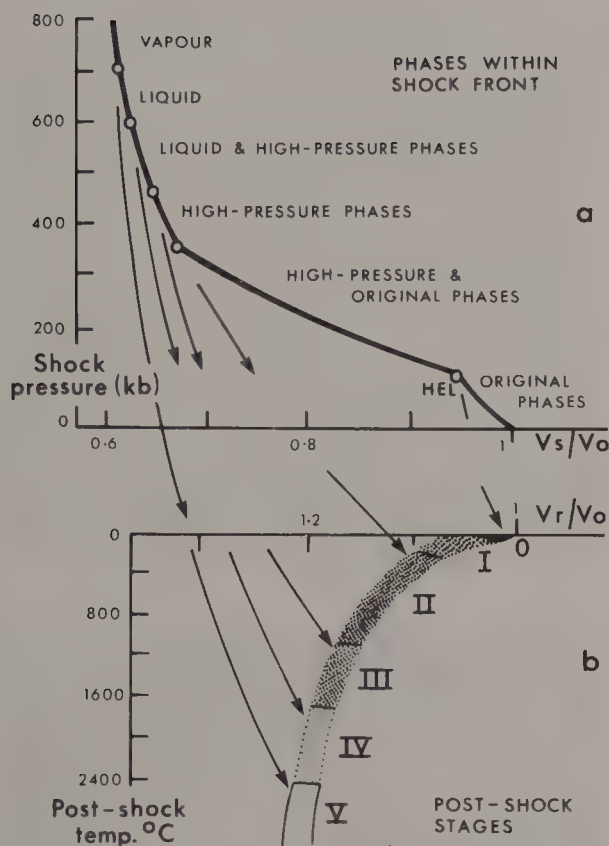


FIGURE 1. Pressure-volume-energy states in shocked material. (a) Schematic Hugoniot (heavy line) for framework silicates projected in the p - V plane, indicating Hugoniot elastic limit (HEL) and ratio of shocked volume to original volume (V_s/V_o) decreasing with increasing shock pressure; small circles on Hugoniot delimit ranges for phase assemblages within the shock front (10 kb = 1 Gigapascal). (b) Post-shock temperatures vs. volume ratio (V_r/V_o) for phases developed following pressure release; stages O-V are defined by representative post-shock phases (below); their relationships to successive regimes in (a) are indicated by arrows. (After Stöffler, 1972.)

Stage O	quartz and feldspar fractured, kink bands in mica.
Stage I	planar features in quartz and feldspar, partial transformation of feldspar to diaplectic glass.
Stage II	diaplectic SiO_2 and feldspar glass, coesite and stishovite (rare), deformation bands in amphibole and pyroxene.
Stage III	feldspars vesiculated and melted, quartz remains as diaplectic glass; ferromagnesian silicates are decomposed.
Stage IV	inhomogeneous glass from complete melting.
Stage V	homogeneous glass from vaporization.

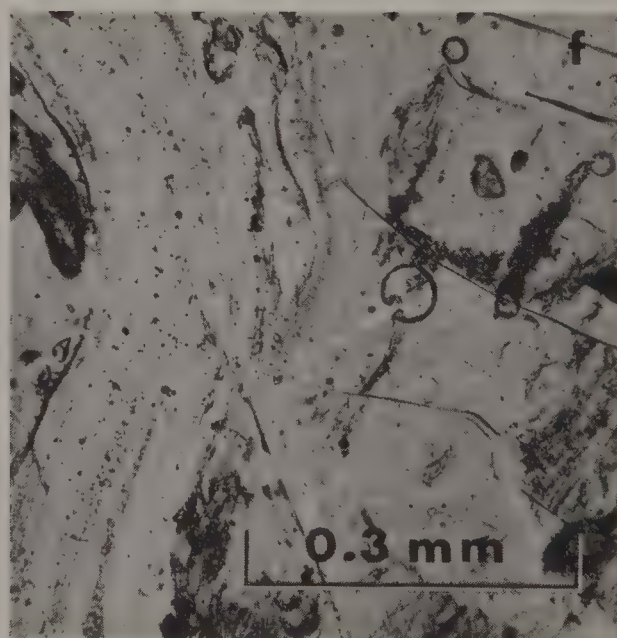
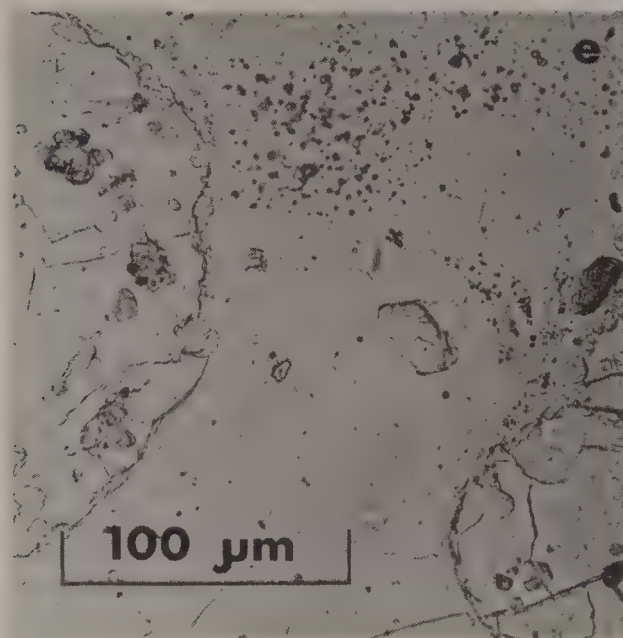
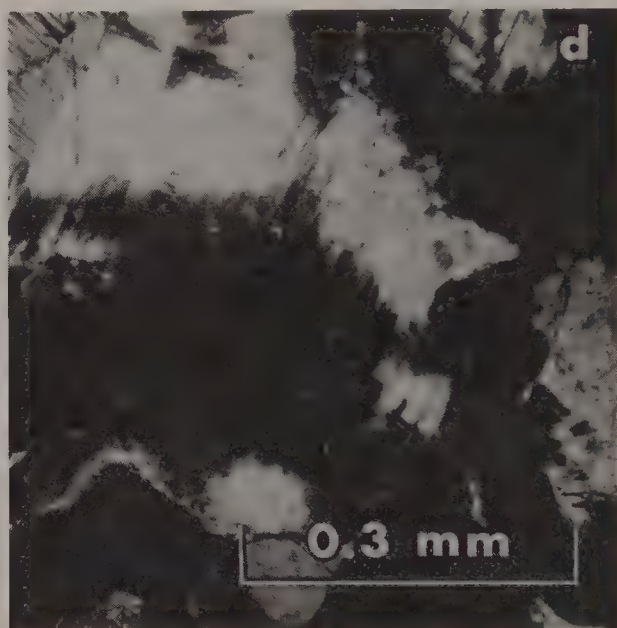
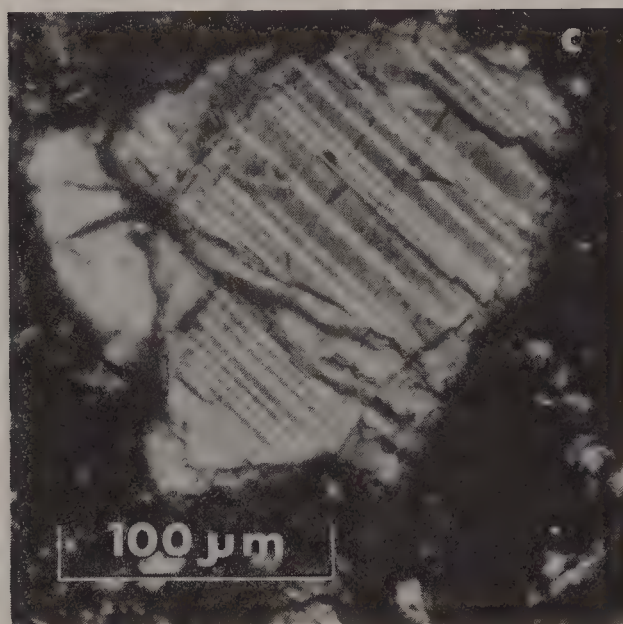
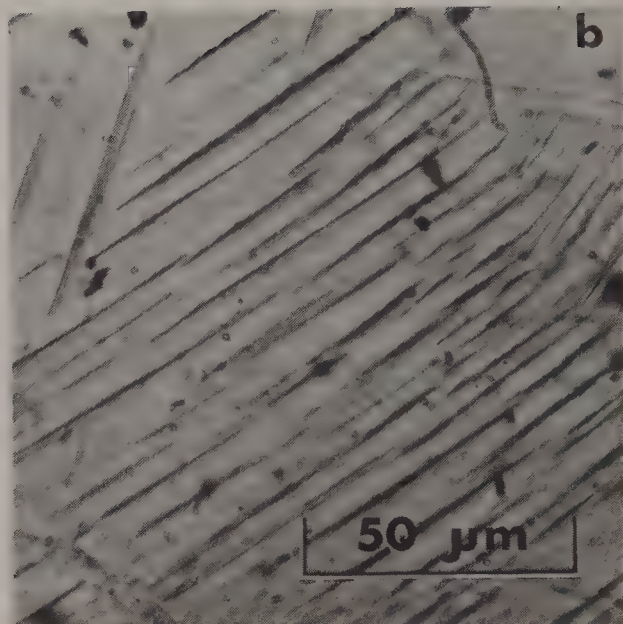
plete. They are best developed in fine-grained rocks and result from shock-wave interactions at pressures generally of a few tens of kilobars (Dietz, 1972).

Rocks produced as mixtures of highly

shocked materials (*impactites*) include *suevite*, a glass-rich breccia first described from the Ries crater, Germany, and large bodies of *impact melt rocks* (Grieve et al., 1977). The melts that result in such glassy or recrystallized igneous-textured bodies are formed by thorough mixing of materials completely melted or vaporized by extreme (>600 kb) shock pressures (*shock melting*).

Most of the more distinctive effects of shock are observed in minerals at the microscopic scale.

1. *Planar fractures*, common in open framework minerals and parallel to crystallographic planes of simple Miller indices, form over a wide range of shock pressures and may be distinguished from fractures produced at low strain rates by their frequency distribution.
2. *Planar features or elements* like multiple sets of closely spaced optical discontinuities or narrow lamellae are produced above the HEL in quartz, feldspar, olivine, sillimanite, and other minerals. In quartz (Fig. 2b), they appear as sets of fine, straight planes 1–20 μm apart, oriented parallel to crystallographic planes, of which (10 $\bar{1}$ 3), (10 $\bar{1}$ 2), (0001), and (10 $\bar{1}$ 1) are the most common. Many are “decorated” with minute cavities or inclusions, largely as the result of later annealing. Other planar elements, particularly in feldspars, take the form of broader lamellae with slightly differing optical orientation from the host. Planar elements are most common in open framework silicate minerals shocked to 100–300 kb, and the average number of sets per grain and the number of sets per unit volume increase with pressure. The mean refractive index and birefringence of these minerals show a concomitant decrease (Stöffler, 1974). Such grains are mixtures of the original mineral plus short-range-order phases, the latter formed within the planar elements by the inversion of dense polymorphs on pressure relaxation.
3. *Deformation bands*, such as kink bands, mechanical twins, and extinction bands with diffuse boundaries, are common in sheet and chain silicates and calcite (Fig. 2c).
4. *Mosaicism*, the general disorientation of crystal structure domains or blocks, best observed using X-rays, accompanies the effects outlined above in the higher end of the shock-pressure range.
5. *Diaplectic* (= “formed by striking”) glasses, including those called *thetomorphic* (= “adopted form”) glasses, are isotropic, short-range-order phases, show no diffraction lines under X-ray, formed in the solid state



on decompression of framework silicates from 250–500 kb. Maskelynite is diaplectic plagioclase glass (Fig. 2d). These glasses have properties distinctly different from those of quenched-fluid glasses (Stöffler, 1974).

6. *High-pressure polymorphs*, such as coesite and stishovite derived from quartz in terrestrial impact structures (Fig. 2e), diamond, and lonsdaleite derived from graphite, and ringwoodite and majorite from olivine and enstatite, respectively, occur in meteorites.
7. *Products of thermal decomposition and melting* occur where postshock temperatures exceeded either the decomposition or the melting temperature of minerals. Biotite, amphibole, and zircons decompose, and framework silicates soften or melt at pressures above ca. 450 kb. In granular materials or rocks with appreciable porosity, the heat generated by the collapse of pore space induces partial or complete melting at significantly lower pressures (Schaal et al., 1979).

Stages of shock metamorphism

By using experimental results to calibrate the mineralogical changes, a scheme of generalized stages of progressive shock metamorphism has been proposed for quartzofeldspathic rocks of negligible porosity (Fig. 1b). Shock zones at impact sites have been mapped using this scheme (Robertson, 1975) and have been used to analyse apparent shock-wave attenuation and deformation in natural impact craters (Dence et al., 1977).

Other terms

Complex crater—a meteorite impact crater of large diameter and relatively shallow depth; a

central uplift and peripheral ring depression developed in the late stages of the crater-forming event.

Kärnäite—an agglomerate-like tuff with a glassy groundmass and a composition similar to dacite that was originally described as a volcanic rock; subsequently interpreted as an impactite with bedrock fragments (Kärnä Island, Lake Lappajarvi, Finland).

Köfelsite—a pumiceous high-SiO₂ glass containing shock-metamorphosed mineral fragments and shock-melted glass that occurs in small veins and is interpreted as representing the results of meteorite impact (Köfels, Austria).

Primary crater—an impact crater produced by the high-velocity impact of a meteorite or other projectile.

Secondary crater (satellitic crater)—an impact crater produced by the relative low-velocity impact of fragments ejected from a primary crater.

Shock lithification—instant rock formation by fracturing, compaction, and intergranular melting by the action of shock waves; originally loose fragmental materials converted into coherent aggregates.

Shock-metamorphic facies—the term for an association of mineralogical features (e.g., multiple parallel planes and thetomorphic glasses) formed by a particular degree of metamorphism; not generally used, since a near approach to equilibrium is implied.

Simple crater—a meteorite impact crater of relatively small diameter, a uniformly concave-upward shape, a maximum depth in the center, and lacking a central uplift.

Thetomorph (thetomorphic glass)—glasses or glassy phases produced by solid-state alteration of an originally crystalline material by the action of shock waves, commonly of quartz or

FIGURE 2. Features of shock metamorphism.

- (a) Shatter cones in quartzite; Kelley Lake, Sudbury, Ontario, Canada. (Photograph by R. Dietz.)
- (b) Sets of planar features with {10 $\bar{1}$ 3} orientations in quartz; Hardhat nuclear explosion, Nevada Test Site, U.S.A.; PPL.
- (c) Deformation bands in clinopyroxene; lunar breccia 14314; XN. (From *Proc. 3rd Lunar Sci. Conf.* 1, 1972.)
- (d) Partial transformation of plagioclase (An₅₄) to maskelynite; Mont de Babel, Manicouagan crater, Quebec, Canada; XN.
- (e) Small clusters of high refractive index coesite in two grains of diaplectic silica glass (left and lower right), separated by maskelynite containing a small apatite crystal; Zipplinger, Ries crater, W Germany; PPL.
- (f) Vesicular feldspar glass showing flow lines (left) in a matrix of homogenized melt containing many small inclusions; West Lac à L'Eau Claire crater (Clearwater Lake), Quebec, Canada; PPL.

feldspar, and retaining their forms and original textures.

M. R. DENCE
P. B. ROBERTSON

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Cross-references: *Astroleme*; *Cataclastic rocks*; *Cryptovolcanic structure*; *Lunar petrology*; *Meteorites—petrology*.

SILICATE MELTING

Nature of the silicate chemical bond

The silicates are predominately ionic compounds. In ionic compounds, as a first

approximation, the ions are assumed to be rigid spheres. The mutual potential energy (V_{ij}) between ions is then the sum of the Coulomb attraction and a repulsive energy which is the result of electronic interpenetration. The equilibrium distance between the centers of charge, or the radius sum ($R_i + R_j$), is that separation which will minimize the mutual potential energy according to the relationship

$$B_{ij} = \frac{z_i z_j e^2}{(R_i + R_j)} + \frac{b_{ij} e^2}{(R_i + R_j)^n} \quad (1)$$

where z_i and z_j are the respective charges of the cation and anion and b_{ij} is the repulsion coefficient. The total energy of a crystal can be obtained by summing up the terms V_{ij} over all pairs of ions in a crystal.

The ions in a crystal are in constant agitation. As the temperature increases the agitation becomes violent enough to destroy long-range ordering and rigidity and as a consequence melting results. In ionic compounds melting takes place when kt (the Boltzmann constant $\times$ the absolute temperature) is equal to about 1/10 energy of the weakest crystal bond. However, in practice, V_{ij} gives only a crude approximation to the bond energy because ions are deformable (polarizable) and a more refined approximation must take this into consideration.

The silicates are classed as ionic mesodesmic structures because the fundamental unit in silicates is the silica tetrahedron in which 1/2 the charge of the oxygen anions (z_o) is satisfied by the tetrahedral link. The remainder of the charge is available for network formation (polymerization). The Si^{4+} cation, which has a high ionic potential ($z_i/R_i = 10$), is classed as a network former. Cations with a low ionic potential (<2.7), which do not form networks but fill the interstices in them, are called network modifiers. These cations provide the weakest link in the silicate structure and have a dominant influence on the melting temperature.

Both network formers and modifiers fall short of the rigid ionic sphere abstraction, and as deformation increases the covalent contribution to the bond energy also increases. For example, the Si–O bond has been estimated to be 55% ionic and 45% covalent. According to Pauling (1960) the total diatomic bond energy ($D(A-B)$) is equal to the geometrical mean of the covalent bond energies ($D(A-A)$ and $D(B-B)$) plus the extra ionic contribution, or

$$D(A-B) = (D(A-A)D(B-B))^{1/2} + \Delta \quad (2)$$

In general, for ionic compounds, Δ is dominant and the covalent term does not vary much for diadochic elements. It can be shown

that the melting temperature of silicate structures can be raised by the solid solution of cations which maximize the extra ionic contribution Δ.

Thermodynamic considerations

From the thermodynamic viewpoint the Gibbs free energy difference between the liquid and solid state at the melting temperature is zero:

ΔG = ΔH_f - T ΔS_f = 0 (3a)

or

ΔH_f = T ΔS_f (3b)

where ΔH_f is the enthalpy of fusion and ΔS_f is the entropy of fusion. ΔH_f represents the work required to overcome attractive forces and ΔS_f is a measure of the decrease in ordering upon melting. Values for these functions are given for major rock-forming minerals in Table 1.

TABLE 1. Enthalpy and entropy of fusion for major rock-forming minerals

Mineral	Melting T K	ΔH _f Cal/gm	ΔS _f Cal/gm dep. ×10 ³	ΔT/ΔP °C/K bar
Forsterite	2,163	109	50	4.5
Fayalite	1,478	71	48	3.9
Enstatite	1,830*	147	80	15.0
Diopside	1,664	102	61	13.0
Anorthite	1,823	105	58	3.6
Albite	1,391	49	35	24.7
Orthoclase	1,423*	100	70	12.4
Leucite	1,959	31	16	17.9
Quartz	1,743	58	33	49.0

* Incongruent melting.

Melting of silicates and other ionic structures

Pauling (1960) has shown that the ratio of the experimental electric dipole moment to the electric dipole moment calculated for a rigid ionic structure is given by the following relationship

μ_{exp} / μ_{rigid} = (1 - e^{-0.25Δ²}). (4)

He refers to this ratio as the fractional ionic contribution. If it is assumed that the bond length remains constant, then the lower values of μ_{exp} relative to μ_{rigid} can be ascribed to the transfer of a fraction of the formal charge from cation to anion. If *e* is the formal charge and *e'* is the effective charge, then

F = $\frac{e'R}{eR} = \frac{e'}{e} = (1 - e^{-0.25\Delta^2})$ (5)

and consequently,

e' = *Fe* (6)

where *F* is, by definition, identical to the expression on the right hand side of Eqn. (4) as indicated in Eqn. (5).

The single bond ionic resonance energy can then be calculated from the equation

V₁ = $\frac{F^2e^2}{R} = \frac{14.40F^2}{R}$ (7)

where 14.40 is the ionic bonding energy in electron-volts (e.v.) of a rigid molecule with bond length of 1 Å. But *F* is the fractional ionic contribution and, so, dividing Eqn. (7) by *F*, the total bond energy (*V_T*) of a diatomic bond involving a single electron is obtained. If the cation has a nominal charge *Z_c* in coordination with *N_c* anions with charge *Z_a* and the bond length is the radius sum (*R_a* + *R_c*), then the bond energy for a cation-anion pair is

V_T = $\frac{14.40FZ_aZ_c}{N_c(R_a + R_c)} + \frac{V_s}{N_c}$ (8)

where the term *V_s*, the crystal field stabilization energy, must be added for transition elements.

The effect of these variables on melting temperature can be demonstrated by reference to minerals with the NaCl (halite) structure. For example, the alkali fluorides demonstrate the effect of radius sum:

	LiF	NaF	KF	RbF	CsF
(<i>R_a</i> + <i>R_c</i>) Å	2.04	2.33	2.69	2.83	3.03
<i>F</i>	0.89	0.91	0.92	0.92	0.93
<i>V_T</i> (e.v.)	1.02	0.93	0.82	0.78	0.74
M.P. °C	870	980	860	780	680

The effect of electronegativity on alkali fluoride pairs with the same radius sum can be seen by the following halite structure pairs:

	LiBr	KF	LiI	RbF
(<i>R_a</i> + <i>R_c</i>) Å	2.63	2.69	2.84	2.83
<i>F</i>	0.56	0.92	0.47	0.92
<i>V_T</i> (e.v.)	0.51	0.82	0.40	0.78
Melting point °C	550	860	450	780

The combined effect of increasing radius sum and decreasing *F* can be observed for the lithium halides (halite structure):

	LiF	LiCl	LiBr	LiI
(<i>R_a</i> + <i>R_c</i>) Å	2.04	2.49	2.63	2.84
<i>F</i>	0.89	0.63	0.56	0.47
<i>V_T</i> (e.v.)	1.02	0.61	0.51	0.40
Melting point °C	870	610	550	450

The effect of increased charge is evident in the high melting point of the alkaline earth

oxides which also have the halite structure:

	BaO	SrO	CaO	MgO
$(R_a + R_c)$ Å	2.74	2.52	2.39	2.06
F	0.83	0.79	0.79	0.73
V_T (e.v.)	2.91	3.01	3.17	3.40
M.P. °C	1920	2430	2570	2800

For ionic compounds with the fluorite structure and $Z_c = 2$, $Z_a = 1$, and $N_c = 8$, the data in Table 2 are obtained.

The influence of structure can be seen by the sudden change in melting point between AlF_3 and SiF_4 in the following series:

	NaF	MgF ₂	AlF ₃	SiF ₄	PF ₅	SF ₆
M.P. °C	980	1400	1040	-77	-94	-56

The fluorides of Na, Mg and Al form structures that are composed of infinite 3-dimensional networks of ions, whereas the fluorides of Si, P, and S form molecular crystals. This is a result of the saturation of charge in the SiF_4 , PF_5 , and SF_6 neutral complexes. For example, silica prefers a 4-fold coordination and the 4 fluorine anions completely satisfy its charge leaving only van der Waal's forces remaining to form an infinite network. On the other hand, the preferred coordination of six F^- around Al^{+3} requires a sharing of anions in an infinite network to maintain charge neutrality. The weak van der Waal's bonds are stable only at very low temperatures.

The fact that V_T can be used to predict melting points for the more complex silicates can be shown by reference to the olivine solid solution series, forsterite to fayalite ($N_c = 6$) (Table 3).

A similar relationship is observed for the substitution of Ca^{+2} for Na^{+1} in the plagioclase feldspar solid solution series:

Mineral	Cation	Radius sum (Å)	F	V_T (e.v.)	M.P. (°C)
Anorthite	Ca^{2+}	2.39	0.79	3.2	1550
Albite	Na^{+1}	2.37	0.82	1.7	1118

In this case, although the Na^{+1} vs. Ca^{+2} substitution, which is the weakest link with oxygen in the silicate structure, dominates the melting point, the actual substitution is $Na^{+1}-Si^{4+}$ vs. $Ca^{2+}-Al^{3+}$, and it can be shown that the concomitant Si^{4+} vs. Al^{3+} substitution has an influence of secondary importance. For example, consider substitution in albite and orthoclase. The values are as follows:

Cation	Ligancy	V_T (e.v.)
Si^{4+}	4	8.3
Ge^{4+}	4	7.4
Al^{3+}	4	7.1
Ga^{3+}	4	6.4
Al^{3+}	6	4.7
Ga^{3+}	6	4.3

The melting points of artificial albite ($NaAlSi_3O_8$) as given by Goldsmith (1950) are:

Mineral	M.P. (°C)	Substitution	ΔV_T (e.v.)
Albite	1118	none	0.0
Ge-albite	1067	$Ge^{4+} \rightarrow Si^{4+}$	-0.9
Ga-albite	1015	$Ga^{3+} \rightarrow Al^{3+}$	-0.7
Ge-Ga-albite	952	$Ge^{4+}Ga^{3+} \rightarrow Si^{4+}Al^{3+}$	-1.6

Thus, the substitution of lower V_T cations lowers the melting point step by step. Furthermore, this is also evident for orthoclase ($KAlSi_3O_8$):

Mineral	M.P. (°C)
Orthoclase	1170
Ge-orthoclase	1122
Ga-orthoclase	1020
Ge-Ga-orthoclase	1000

The influence of pressure has been neglected thus far. Silicates expand upon melting and so pressure tends to stabilize the mineral according to Le Chatelier's principle and the Clausius-

TABLE 2. Data for ionic compounds with fluorite structure

	CaF ₂	SrF ₂	BaF ₂	CdF ₂	PbF ₂	HgF ₂
$(R_a + R_c)$ Å	2.35	2.49	2.71	2.33	2.56	2.46
F	0.90	0.90	0.91	0.73	0.70	0.67
V_T (e.v.)	1.37	1.30	1.21	1.13	1.01	0.98
Melting point °C	1420	1400	1320	1110	820	650

TABLE 3. Data for forsterite and fayalite

Mineral	Cation	$(R_a + R_c)$ (Å)	F	V_s/N_c (e.v.)	V_T	M.P. (°C)
Forsterite	Mg^{2+}	2.06	0.73	0.00	3.40	1905
Fayalite	Fe^{2+}	2.14	0.56	0.09	2.60	1205

Clapeyron relationship

$$\frac{\Delta T}{\Delta P} = \frac{T \Delta V}{\Delta H} = \frac{\Delta V}{\Delta S} \quad (9)$$

$\Delta T/\Delta P$ values for common silicate minerals are given in Table 1.

Melting of binary silicate systems

The type of phase diagram obtained with binary silicate mixtures depends primarily on the radii of cations present within the end members, providing that charge balance may be maintained in intermediate mixtures. If the radii are similar in size, as in the olivine series, a continuous ascending solid solution will be obtained. When the cation radii are dissimilar ($>25\% < 60\%$), there is a tendency to form solid solutions with minima. Albite ($\text{NaAlSi}_3\text{O}_8$)–orthoclase (KAlSi_3O_8) is a familiar example of a binary solid solution with a minimum [$R_c(\text{Na}^{+1}) = 0.97 \text{ \AA}$, $R_c(\text{K}^{+1}) = 1.33 \text{ \AA}$]. When there is a charge imbalance or the cation radii differ greatly ($>60\%$), the end members will be structurally different and a eutectic minimum will be obtained, e.g., anorthite ($\text{CaAl}_2\text{Si}_2\text{O}_8$)–diopside ($\text{CaMgSi}_2\text{O}_6$).

In ascending solid solutions the melting point of mixture is intermediate between that of the end members. In binary systems with a minimum, the freezing depression is given approximately by the equation for ideal concentration solutions

$$T = \frac{T_0}{1 - \frac{RT_0 \ln x}{\Delta H_f}} \quad (10)$$

where ΔH_f is the molar latent heat of melting and x is the mole fraction of the end member whose melting temperature is T_0 . Where appreciable differential heats of mixing are involved, the observed melting point will diverge from the simple theoretical relationships.

Melting temperature of silicate rocks

Silicate rocks being diverse mixtures will tend to melt at a lower temperature than the constituent minerals. Furthermore, the addition of a network modifier to a silicate melt, e.g., Na_2O to water glass (Na_2SiO_3), will impede polymerization and thus further lower the melting point.

There should be an inverse relationship between the silica content and melting point of dry rocks since it is the polymerization of silica tetrahedra which forms the basis of most silicates (neosilicates, e.g., olivine, excepted) and the complexity and susceptibility to

disordering of the network increases with silica content. That this relationship does exist is well known and can be demonstrated by the following data:

Type of rock	Wt. % SiO ₂	Complete melting (°C)
Meteoritic silicate	45	1450
Gabbro	49	1250
Diorite	57	1090
Quartz diorite	62	1000
Granodiorite	65	930
Granite	70	830
Pegmatite	72	750

Addition of singly charged anions such as OH^- (H_2O), F^- or Cl^- to a melt will also decrease polymerization of the melt and profoundly lower the melting temperature. Such anions will have their charge satisfied by a single Si^{4+} cation and thus the need for further linkage is eliminated. The effect of water pressure far exceeds the opposing dry pressure effect. For example, 5 kb $\text{P}_{\text{H}_2\text{O}}$ lowers the melting point of granite by 400°C . This disordering and mobilizing effect of water and other suppliers of singly charged anions is of exceedingly great importance in many geological processes including the formation of ore deposits as well as the crystallization of magmas.

Other terms

Congruent melting—melting of a substance directly to a liquid that is of the same composition as the solid.

Incongruent melting—melting to give a liquid different in composition from the original solid.

P. E. DAMON

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Cross-references: *Anatexis; Magma*.

SILICATES—HYDROTHERMAL SYSTEMS—See Vol. IVA

SILL

A *sill* is a planar body of igneous rock intruded parallel to bedding or other primary structure of its host rocks, i.e., a concordant intrusion. Its thickness is very constant (Fig. 1) and is generally of the order of one hundredth of the lineal extent. Laterally, sills terminate abruptly against faults or finger out as one or several interdigitating wedges. In rare cases they pass laterally into curved, inclined sheets (basin sheets) and then into dykes (Lombaard, 1952). Sill contacts with enclosing rocks are generally regular and sharp and there may be no admixture of the rocks over hundreds of square kilometers. In a few sills the contacts are locally complex and brecciation of wall rocks is intense. Thermal metamorphism of contact rocks is characteristically slight, but in rare instances roof rocks are fused or metasomatized.

Sills fall into two contrasting categories of size, composition, and tectonic environment (Bradley, 1965). Major sills are regional structures commonly exceeding 1000 km² in area, 300 m in thickness, and 100 km³ in volume. They belong to stable areas of the continental platform and are intruded at relatively shallow depths (Mudge, 1968) into horizontal and shallow water sediments, usually deltaic. They consist exclusively of dolerite or norite, and at the time of their emplacement the sediments they intrude were almost always very young, relatively weak, and of low density.

These intrusions are regarded by Bradley as the result of flotation intrusion in the absence of high differential pressure (but see Gretener (1969) and Roberts (1970) for alternative views and references).

Minor sills comprise the great majority by number and type of this form of intrusion. Typically they belong to a localized volcanic environment and are usually attached to individual volcanoes. They are generally less than 100 km² in extent and 5 km³ in volume. They include rocks of all compositions, including dolerite, and are intruded into the lavas, tuffs, and sediments around the flanks of volcanoes. Minor sills are probably intruded under high pressures of magma in the volcanic vent, and are similar in this regard to laccoliths.

All sills are emplaced as fluid or partly crystallized magmas that cool and solidify over periods of three days for sills a meter thick to tens of years for major sills (Jaeger, 1958). Thin sills tend to be finer grained than thicker ones and all sills tend to be finer towards their margins. Acidic rocks, in particular, often solidify largely as glass or obsidian.

Sills, the larger ones in particular, are probably intruded in pulses at intervals of several years. This gives rise to several types.

Simple sill—a uniform sill intruded as a single pulse.

Multiple sill—a series of sills of similar composition intruded next to, or within, each other; units are distinguished by fine-grained chilled contacts of the later against earlier sills.

Composite sill—usually consists of two or three sills of different composition intruded

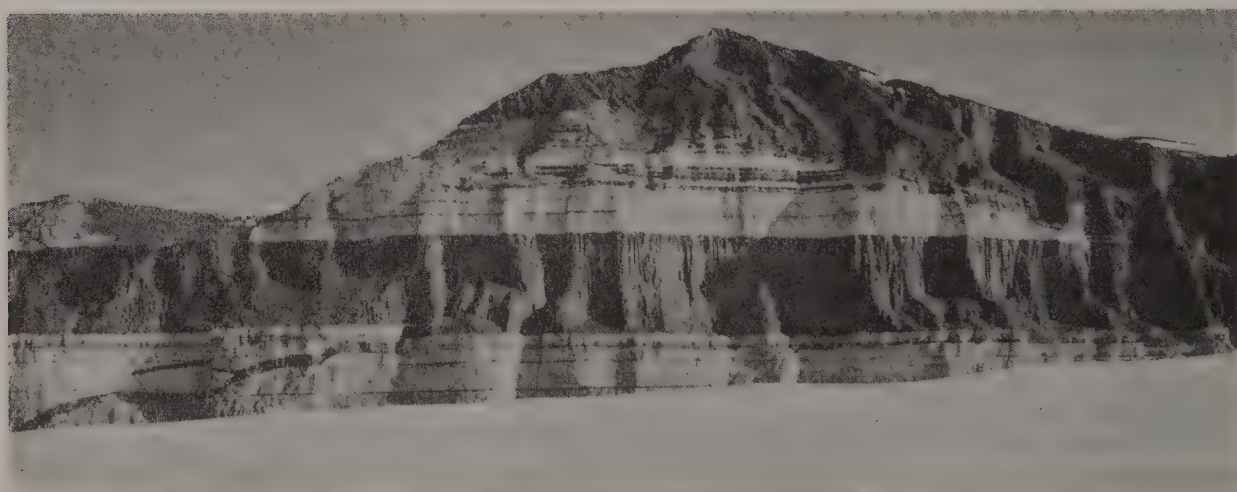


FIGURE 1. Dolerite sill (black) in Beacon Sandstone, Taylor Glacier, Victoria Land, Antarctica; also shown are several thin, simple sills (center), a large inclined sheet (top right), and inclined sheets passing into sills (bottom left).

within each other; the most common association is of granophyre within dolerite.

Differentiated sill—stratified in layers of varied composition owing to the differential settling of early-formed and basic minerals; the process is complicated by the pulsative nature of intrusion and, in thick sills, by convective sorting of minerals; differentiated major sills are transitional into lopoliths.

JOHN BRADLEY

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Cross-references: *Dolerite*; *Dyke (dike)*; *Emplacement—modes*; *Hypabyssal rocks*; *Igneous intrusions—structural behavior*; *Laccolith*; *Phacolith*.

SKAERGAARD INTRUSION

The first account of the *Skaergaard intrusion* in E Greenland (Wager and Deer, 1939) was based on the results of field work in 1935–1936

and subsequent laboratory studies. After the Second World War this “Skaergaard memoir” quickly became the “avant garde” of much igneous petrologic thinking, and the study of layered intrusions, the cumulates they contain, and the rhythmic and cryptic layering they exhibit, became a major field of research. A follow-up expedition to E Greenland was not possible until 1953. Its results and a review of the enormous expansion of information relating to layered (stratiform) intrusions was published by Wager and Brown (1968).

The intrusion is funnel-shaped and dominated by gabbroic rocks. It is of Eocene age and was intruded into Precambrian gneisses and Tertiary basalts (Fig. 1), being subsequently tilted during the development of the Greenland coastal flexure. Its oval-shaped outcrop occupies an area approximately 9×7 km. Apart from some cover by the glaciers (much diminished in volume since 1936), there is almost complete exposure (Fig. 2). The intrusion was divided initially into three main parts (Wager and Deer, 1939): a Marginal Border Group, an Upper Border Group, and a Layered Series, the latter forming the bulk of the intrusion. The Layered Series was later further divided into Lower, Middle, and Upper zones, with a Hidden Zone below (Wager and Brown, 1968; McBirney, 1984).

The *Marginal Border Group* is dominantly picritic, and is presumed to have formed by the accumulation of early-formed olivine crystals. Elsewhere it shows marginal chilling, succeeded inwards by wispy banding, due to varying proportions of light and dark minerals. In one part is developed an unusual perpendicular feldspar rock, where long, branching feldspar crystals extend inwards at right angles to the margin of the intrusion.

The most rewarding study within the intrusion has been that of the *Layered Series*, a structurally continuous succession of layered

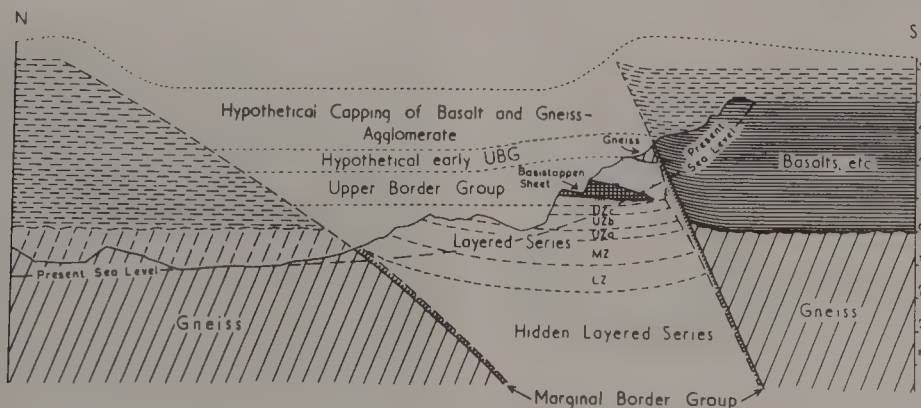


FIGURE 1. Schematic section of the Skaergaard intrusion before monoclinial flexuring and erosion. (From Wager and Brown, 1968.)



FIGURE 2. Northern part of the Skaergaard intrusion, looking east. (From Wager and Brown, 1968.)

gabbros some 2,500 m in thickness. The layering is frequently rhythmic (Fig. 3), homogeneous gabbro giving way upwards quite sharply into a melanocratic band rich in ferromagnesian minerals that grades up into a feldspar-rich unit, itself overlain again by normal homogeneous gabbro. The fact that everywhere less-dense feldspathic bands overlie denser melanocratic bands suggests that gravity has played an important role in the formation of these rocks. A pronounced parallelism of the platy plagioclase and pyroxene crystals, producing igneous lamination, is considered evidence that magmatic currents have been operative. The component minerals exhibit cryptic (or hidden) variation (from the base of the Layered Series to the top):

plagioclase—basic labradorite (An_{69}) to andesine (An_{34}),
 olivine—chrysolite ($Fo_{69.5}$) to fayalite ($Fo_{2.5}$),
 and
 clinopyroxene—augite ($Fe_{9.7}$) to ferroaugite ($Fe_{34.8}$).



FIGURE 3. Rhythmic layering with gravity-stratified layers separated by layers of homogeneous gabbro; Skaergaard intrusion. (From Wager and Brown, 1968.)

The orthopyroxene shows a similar Fe enrichment.

These changes in composition, which are from higher- to lower-temperature forms, led Wager and Deer to deduce that the order of formation of the Layered Series was from below upwards, i.e., the bulk of the Skaergaard intrusion formed by bottom accumulation of primary precipitate (cumulus) crystals. The production of exceptionally Fe-rich ferrogabbros from a tholeiitic magma under conditions of strong fractionation was considered to be the normal differentiation trend for this type of magma; it is often referred to as the "Skaergaard trend."

In the higher parts of the Layered Series, the even layering is interrupted by a different type of banding, in the form of long, narrow troughs (Fig. 4). About twenty such troughs have been located, and although they occur at different structural levels in the Series, all tend to radiate from the center of the intrusion. They exhibit pronounced banding, with noticeable concentration of heavy minerals (particularly Fe-ore and olivine) towards the center of each trough. The dips of the troughs are generally similar to those of the adjacent layering.

The origin of the various features present in the Skaergaard intrusion was ascribed by Wager and Deer to convection currents, instigated by denser crystal plus liquid near the margins of the intrusion moving downwards and across the floor towards the center, depositing the crystals as a mush as they move. Compensating currents move upwards in the center of the intrusion (see Wager and Brown, 1968). Differential gravitational settling of the heavy, dark minerals was considered responsible for the rhythmic layering present in the Layered Series and the igneous lamination was interpreted as representing the orientation of platy minerals by the magmatic currents. The trough banding was explained by the splitting of convection currents, into narrow, powerful streams, producing an even more



FIGURE 4. Trough banding; Skaergaard intrusion. (From Wager and Brown, 1968.)

pronounced separation of light and dark minerals (cf., Irvine, 1987).

The *Upper Border Group* was considered to have formed by solidification of successive residual magmas (without primary crystals) downward from the top; there is a downward cryptic variation comparable with the upward cryptic variation in the Layered Series.

Drilling through much of the Skaergaard intrusion subsequently led to petrologic concepts such as the tendency for liquid in a magma to undergo compositional thermal and density stratification prior to solidification, and an enormous expansion in knowledge of basaltic magma has resulted in modification of some of the original postulates of Wager and Deer (cf., McBirney and Noyes, 1979; Irvine, 1982; McBirney, 1984—see **Intrusion—layered**). However, it could be argued that the Skaergaard intrusion has had more impact on the evolution of igneous petrology than any other single igneous intrusion.

D. S. WEEDON

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Cross-references: *Basalt*; *Bushveld Complex*; *Cumulate textures*; *Gabbro and norite*; *Intrusion—layered*; *Magmatic currents and flow*; *Magmatic differentiation*.

SKARN

Skarn is a poorly defined term used in several imprecise ways. It usually refers to a holocrystalline, massive, bedded or banded rock

rich in calc-silicate minerals, which may or may not carry ore minerals. It is frequently regarded as the product of metasomatism, pyrometamorphism, or pyrometasomatism. In the English-speaking world “skarn” is usually reserved for contact metamorphosed limestones or marbles carrying a variety of calc-silicate minerals, namely olivine, members of the humite group, vesuvianite, calcic clinopyroxene, calcic amphibole, epidote, ugrandite garnets, and scapolite. Such rocks may also carry ore minerals and borosilicates like tourmaline or axinite. There is usually some evidence of metasomatic chemical exchange between a carbonate protolith and a pluton (Fig. 1). In N America Fe-rich skarns in the contact zones of calc-alkaline plutons are called *tactites* (see **Granites and mineralization**).

In Scandinavia, where the term originated, “skarn” originally referred to the waste rock associated with metasedimentary iron ores (usually magnetite) and was synonymous with “gangue.” Gradually, it has come to be used for any calc-silicate-rich metamorphic rock with unusual mineralogy irrespective of whether it lies in the contact zone of a pluton or whether it is ore-bearing.

In the Soviet literature, “skarn” usage not only follows that in Scandinavia, but is extended to include many metasomatic rock types,

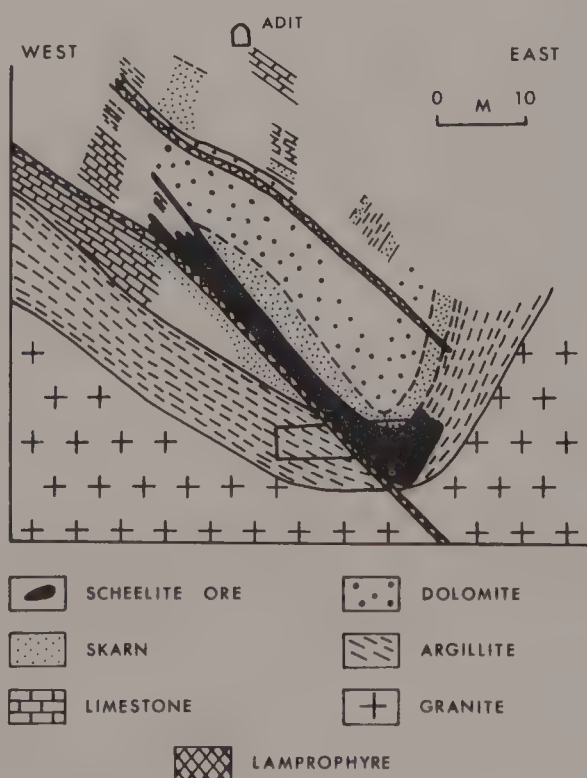


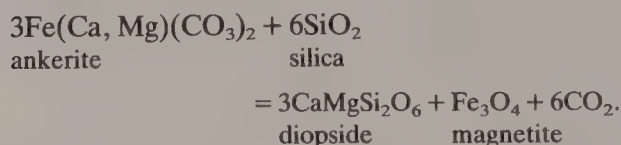
FIGURE 1. Relationship of a scheelite ore body and of skarn to limestone and dolomite adjacent to a granite intrusion; Dodger Mine, British Columbia, Canada. (After Ball, 1954.)

particularly if they carry ore minerals or unusual metamorphic mineral assemblages.

Types of skarn

Most skarns have formed by processes that involved the alteration of a calcareous protolith during or prior to metamorphism. The movement of material within the skarn protolith (redistribution) by hydrothermal action or by massive, pervasive metasomatism can often be demonstrated, although in some cases it can be difficult to demonstrate the introduction of material from without. In some cases skarns may simply represent the metamorphic reconstitution of a protolith that had an exotic composition to begin with. In the simplest cases skarns retain a banding that reflects original compositional layering or bedding. In such cases it is necessary to explain how metasomatism changed the composition of extensive but narrow layers, while failing to homogenize the composition of adjacent layers of widely differing character. While metasomatism remains the preferred explanation for skarn genesis, the subject remains full of contradictions and problems.

Contact skarns. Contact skarns occur in the thermal aureoles of a wide variety of plutonic rocks and invariably involve the contact metamorphism of calcareous sediments (Fig. 1). In the Oslo district Goldschmidt's (1911) classic account of contact metamorphism identified ten mineral assemblages in the high-temperature pyroxene hornfels facies, six of which involved calc-silicate minerals (Fig. 2). Many of the skarn assemblages (essentially Fe-rich variants) can be considered as variants where Fe concentration and oxygen fugacity provide additional variables. If silica is also considered as a variable, then for low a_{SiO_2} assemblages wollastonite would disappear, Ca-plagioclase would be replaced by scapolite or melilite (akermanite), and diopside would be replaced by humite group minerals. By considering such components as Fe and SiO_2 as being mobile or variable, it is possible to describe mineral assemblages as diverse and unusual as those seen in many calc-silicate contact skarns. Similar manipulations may also illustrate the mechanisms whereby metamorphic re-equilibration of carbonate-silicate assemblages produces magnetite, e.g.,



The skarns of the Oslo district are also host to sulfide mineralization, principally sphalerite,

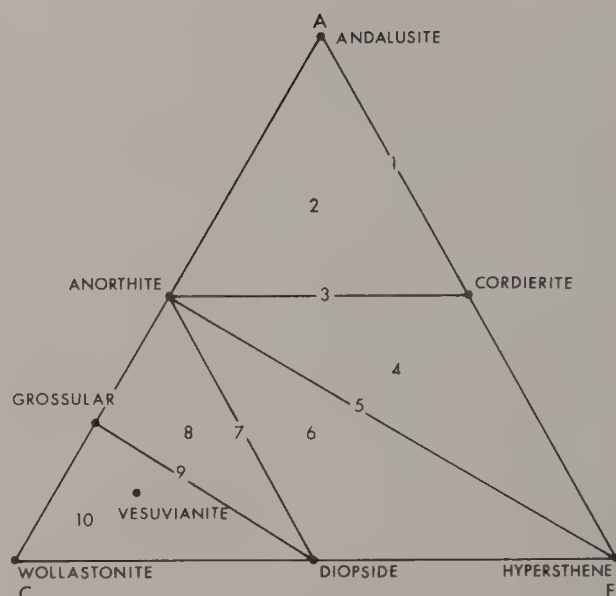


FIGURE 2. ACF diagram of mineral assemblages in pyroxene hornfels facies rocks, Oslo district, Norway. Possible skarn assemblages are (Fe-rich in brackets): 7, plagioclase–diopside (plagioclase–hedenbergite); 8, plagioclase–diopside–grossular (plagioclase–hedenbergite–grossular or andradite); 9, diopside–grossular ± vesuvianite (hedenbergite–grossular or andradite ± vesuvianite); 10, diopside–grossular–wollastonite ± vesuvianite (hedenbergite–grossular or andradite–wollastonite ± vesuvianite). (After Goldschmidt, 1911.)

pyrite, and galena, occurring in veins. Quartz as a major gangue mineral implies considerable silica migration, and secular considerations suggest it to be coeval with skarn formation.

In porphyry-copper deposits, skarn formation appears to be a function of the interaction of hydrothermal brines with limestone layers in the country rock. Such skarns are commonplace where the mineralized aureole to the calc-alkaline pluton is developed in sedimentary sequences including carbonates, and occasionally these deposits are economically important sources of Cu, Mo, Au, Bi, and Te. Effectively, such skarns represent calcareous hornfels into which hydrothermal solutions have introduced particularly effectively the metals circulating in the convecting cell (see **Granites and mineralization**). Sn-bearing skarns in the aureoles of granites in SE Asia may represent a similar phenomenon, where hydrothermal mineralization is superimposed on the effect of thermal metamorphism of limestone and calcareous shales (Hutchison, 1983). In such complex deposits three processes operate either in concert or during an evolving sequence: (1) thermal metamorphism of carbonates and calcareous shales, (2) hydrothermal introduction of metal-rich brines, and (3) movement of silica. The first produces a calc-silicate “gangue” or

matrix, that may include Fe or Mn oxide and silicate minerals. The second process produces pervasive or vein-fracture-controlled sulfide or oxide mineralization, in which the nature of the brines influences the presence of such minerals as fluorite, carbonates, silica, or chlorides. The third may be responsible for the massive zonation in which desilicified-carbonated, decarbonated-silicified, and massive secondary silica precipitation (Mexican onyx) zones can be identified.

In the same way that cupriferous skarns within porphyry-copper mineralizations can be seen as variants on a more common theme, other intrusion-related processes may produce distinctive skarns, e.g., greisenization producing F, Sn, and B skarns, fenitization producing Na-enriched, rare metal, and Cu skarns, and tourmalinization producing borosilicate-rich skarns.

Skarns in regional metamorphic terrane. These are well documented in Scandinavia. The Fe-Mn skarns of the Bergslagen-Grythyttan area of Sweden and the Cr-silicate skarns of Outokumpu, Finland illustrate the variety that exists. Outside Scandinavia, examples may include the scheelite skarns of Austria, the carbonate-hosted stratabound tungsten deposits of Brazil and the Zn-carbonate pyrometamorphic ores of Franklin Furnace, New Jersey, U.S.A. Eskola (1915) grouped such metamorphosed, bedded skarns with other "bedded" rocks of anomalous composition (e.g., anthophyllite-cordierite schists and kinzigite) as examples of regional metasomatism, while Lindgren (1933) considered the ore deposits in this group to be examples of a pyrometamorphic class. At Outokumpu the secular and spatial relationship of Cr-silicate skarn with serpentinite led Eskola (1933) to consider the former to represent a massive reaction zone between the latter and the metasedimentary wall rocks.

Other types. In examples like the Julia Mine, Yenesei Highlands, U.S.S.R. a pure, bedded limestone is cut by a massive skarn, the contact of which has been compared to a "roll-front" type migrating reaction surface similar to groundwater mineralization features seen in many permeable sediments. Similar interpretations, attributing the features of stratabound skarns to premetamorphic alteration or mineralization are offered for many examples. Other skarns, such as the Swedish Fe-Mn deposits and the Outokumpu (Finland) anthophyllite-cordierite rocks may represent nothing more peculiar than the results of regional metamorphism of sediments with unusual composition (Treloar et al., 1981), such as hydrated and leached basic tuffs, Fe-Mn rich

carbonates, or sea-floor stratiform exhalative deposits.

ADRIAN F. PARK

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Cross-references: *Aureole*; *Contact - metasomatism*; *Facies of thermal metamorphism*; *Granites and mineralization*; *Greisen*; *Hydrothermal processes*; *Metasomatism*; *Phase rule*; *Pneumatolysis*; *Pyrometamorphism*.

SLATE

Slate is a fine-grained, pelitic rock characterized by a penetrative planar parting (slaty cleavage) resulting from the almost perfect alignment of phyllosilicates such as muscovite and chlorite. It is often dense and compact, but may be extremely fissile, depending on weathering, structural relations, and composition. The slaty cleavage is independent of bedding and, therefore, may be at any angle to it (Fig. 1). Although the cleavage is the dominant parting surface, bedding may be expressed as laminae or bands resulting from mineralogical or textural differences.



FIGURE 1. Steeply dipping slaty cleavage axial planar to open fold shown by sub-horizontal bedding; Martinsburg Formation, Pennsylvania, U.S.A.

Mineral composition

The mineral composition of slates varies little. Primary mineral constituents are usually aligned muscovite, often in the form of sericite, chlorite, and minute detrital quartz. Feldspar clasts, opaque ores (magnetite or pyrite), graphite, and occasionally carbonates are the most common minor minerals. Variations in composition generally are in the relative proportions of major constituents. Increase in the amount of primary or replacement carbonates results in a calcareous or dolomitic slate.

The color of slates often depends not only on major mineral composition, but on the degree of oxidation of Fe or C. Thus, red and purple slates owe their colour to large amounts of dispersed hematite, green slates to the reduction of hematite, black slates to the presence of graphite or carbon in insoluble residues. Some green slates may also result from a high percentage of chlorite.

Structure

The most prominent feature of slates is the slaty cleavage, which may be defined as a penetrative, planar parting caused by alignment of phyllosilicates normal to the maximum finite compressive strain direction. It is the major fabric element of slate and is penetrative even on a microscopic scale (Fig. 2), thus distinguishing it from other types of cleavage that may later be superimposed. Fine quartz grains, sometimes showing some recrystallization, are usually dimensionally elongated and aligned in the foliation. It is not uncommon to find prominent strain shadows of quartz at the ends of incompressible opaque ores also aligned along the foliation.

A fine streaking, called the long grain, which is visible on the slaty cleavage surface plunging



FIGURE 2. Slaty cleavage defined by mineral alignment of muscovite, chlorite, and elongate quartz; graded bedding shown at bottom of photomicrograph; PPL, $\times 26$.

down the dip of the cleavage parallel to the *a* fabric axis, was thought of by Sander (1911) as the “transport” direction of rock material during deformation. It is currently regarded in terms of strain theory as the “stretching” direction, or direction of maximum elongation.

Origin

Slates derive from the deformation of accumulations of pelitic material, predominantly fine clays such as illite, either as unconsolidated sediments or lithified to a shale. The accumulations are rarely pure clay, and include varying proportions of fine clasts. The pelitic strata may alternate with silt and graywacke beds ranging in thickness from 1 mm to a few meters. When the strata are folded, the minerals become aligned normal to the maximum finite compressive strain direction in response to the deforming compressive stresses. This alignment is parallel to the axial surface of the folds, giving rise to the frequent reference to slaty cleavage as “axial planar cleavage” (Fig. 1).

Inasmuch as there can be only one maximum finite compressive strain direction, it has been recognized that there can only be one true slaty cleavage, despite the number of subsequent deformational structures.

Mechanisms producing mineral alignment

The manner in which minerals assume a preferred orientation to produce the perfect cleavage in slates has long been a matter of concern and controversy. As early as 1856, Sorby succeeded in experimentally producing preferred mineral orientation in a baked iron oxide-clay mixture by causing rotation of the clays under directed stresses; Becker and Day (1905) and Wright (1906) produced prominent preferred orientation by crystallizing minerals in

a stress field. Under these conditions, crystals were elongated parallel to the long axis of the strain ellipsoid, which is the elongation direction, and normal to the compressive strain direction.

Leith (1923) concluded from experimental and microscopic evidence that preferred orientation resulted primarily from recrystallization of consolidated shales, with mechanical rotation playing only a minor role in the process. This was a popular view for a long time for two reasons: (1) microscopic evidence showed clearly that recrystallization did occur in some minerals parallel to the elongation, and (2) it was difficult to envisage a process whereby minerals in a compacted, consolidated rock, such as shale, could break the bonds along grain boundaries and find the space within which to rotate into a new direction.

Two subsequent contributions to the geologic literature led to a re-evaluation and new approach to the study of the mechanisms of slaty cleavage formation. The first of these (Rubey and Hubbert, 1959), described observed occurrences of abnormally high pore-water pressure in rapidly depositing geosynclinal troughs and its attendant result of preventing significant compaction at depths of 2,100 m and more. Following this, Maxwell (1962) described evidence that suggested slaty cleavage in the Ordovician Martinsburg Formation in Pennsylvania, U.S.A., developed while the sediments were still unconsolidated. He concluded that conditions similar to those described by Rubey and Hubbert in contemporary basins had obtained in Ordovician times, allowing phyllosilicates the freedom and space to rotate

mechanically and assume a preferred orientation during deformation. Further studies have added strong support, extended, and provided further details of Maxwell's theory (cf., Alterman, 1973). As a result, mechanical rotation is now regarded by many as a primary mechanism of mineral alignment, at least in those slates where the evidence has been observed. A competing school of thought rejects the rotation hypothesis in favor of pressure solution (Groshong, 1976).

Syntectonic processes. *Mechanical rotation of existing minerals* occurs in unconsolidated sedimentary sequences in which abnormally high pore-water pressures prevent compaction (Fig. 3). Clay minerals and quartz clasts are rotated into stable positions relative to the strain directions. The alignment provided dewatering paths through otherwise impermeable pelites (Alterman, 1973). Continued flattening of the folds, then probably causes volumetric decrease of the pelitic strata, resulting in stress along grain boundaries with consequent pressure solution.

Pressure solution takes place as the result of the solubility of minerals along stressed grain boundaries increasing with increasing pressure. The result is that solution occurs along surfaces normal to the compressive strain direction that are surfaces of greatest strain. Transport and redeposition of the dissolved mineral material then takes place on unstrained surfaces, those in the direction of elongation. This results in minerals elongated normal to the compressive strain direction. This process is probably the primary mechanism of elongation in foliated rocks of higher metamorphic grade than slates; in slates it may be a secondary process in the elongation of quartz, but a primary one in the molecular rearrangement that transforms clay minerals to muscovite. If, however, mechanical rotation of clays in the unconsolidated pelites has occurred, then the recrystallization of clays to mica is a secondary process that utilizes the existing mineral alignment in the deformed sediments and is not the cause of the preferred orientation.

Intragranular gliding due to deformation of individual minerals by slippage along crystallographic planes of relatively weak bonding to produce elongation and alignment in calcite and the micas is a minor process in the formation of slaty cleavage (Knopf and Ingerson, 1938).

Neomineralization is the growth of new minerals from chemicals carried in solution during deformation; it will be influenced by the stress field, causing elongation as growth in the direction of compressive strain is inhibited. This may be a minor process, if at all operative, in

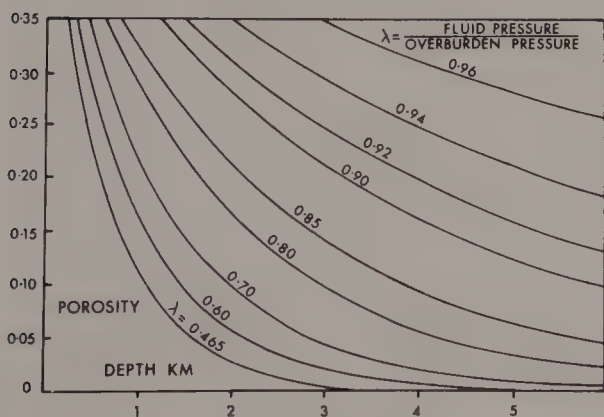


FIGURE 3. Theoretical effects of increasing pore-fluid pressure on compaction; lowest curve ($\lambda = 0.465$) is the normal hydrostatic condition. When fluid pressure within sediment is approximately equal to overburden pressure ($\lambda = 0.96$), porosity at a depth of 5 km is equivalent to porosity of sediment with normal fluid pressure at 0.5 km depth. (After Alterman, 1973.)

the formation of slaty cleavage, as there is little evidence that any of the minerals in slate were not part of the original clastic accumulation, except for muscovite and chlorite that derive from the clay minerals.

Post-tectonic processes. *Mimetic replacement* may take place as the result of the occurrence of thermal highs or conditions favoring retrograde metamorphism following the waning of tectonic stresses. It takes place through solid diffusion or transport by solutions, of existing aligned minerals by new ones that are more stable in the new environment. The new minerals may be pseudomorphic after those replaced, or otherwise utilize the existing alignment. In slates, chlorite may replace biotite in this manner, the post-tectonic chlorite thus assuming the orientation of minerals developed in a stress field. Proof of mimetic growth lies in the fact that the chlorite basal cleavage, which would lie parallel to the mineral alignment if the mineral formed in a stress field, is at high angles to the elongation and foliation.

Where replacement is not a factor, growth of new minerals (neomineralization) in a rock characterized by strong preferred orientation may be restricted by the anisotropy of the rock. While minerals may be able to grow across the foliation, transport of mineral material through solutions or by solid diffusion is most easily accomplished parallel to the foliation, which will influence directions of growth. New minerals may thus assume a preferred orientation without the influence of deforming stresses (Turner, 1981).

Metamorphism

Slate has generally been regarded as the first stage of a progressive metamorphic series that begins with the premetamorphic shale and progresses through slate to phyllite, then mica schist with increase in temperatures, confining pressure, and presence of thermal solutions. However, while this sequence can be traced in some metamorphic regions, there are occurrences of slate completely enclosed by non-crystalline rocks. For example, in the slate belt described by Maxwell (1962), deformation occurred when the pelitic strata were the uppermost units of the sedimentary pile and the slates are presently underlain by highly deformed but noncrystalline carbonates. The further observation that slate in many areas developed directly from unconsolidated sediments leads to reconsideration of slate as a truly metamorphic rock. Since lithification accompanied slaty cleavage formation, partly as a result of the deformation that allowed the dewatering, the process of slate formation lies

somewhere between diagenesis and metamorphism.

Uses

The perfect planar parting of slate and its resistance to chemical weathering have made this rock a commercially useful material. The relatively simple quarrying operations and rapid mechanical means by which minor irregularities of the cleavage surface are removed have resulted in a comparatively inexpensive product, used in schools as blackboards, in buildings as roofing slate, steps, walkways, and table tops, and as a perfect base for billiard tables.

INA B. ALTERMAN

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Cross-references: *Greenschist facies*; *Metamorphic mineral growth*; *Nucleation of metamorphic minerals*; *Phyllite*; *Regional metamorphism*; *Schist*.

SNOWBALL GARNET

A *snowball garnet* (Flett, 1912) is one that contains internal surfaces (visible in thin section as discontinuous lines of inclusions of quartz, magnetite, graphite, etc.) of a characteristic form considered to imply rotation relative to the

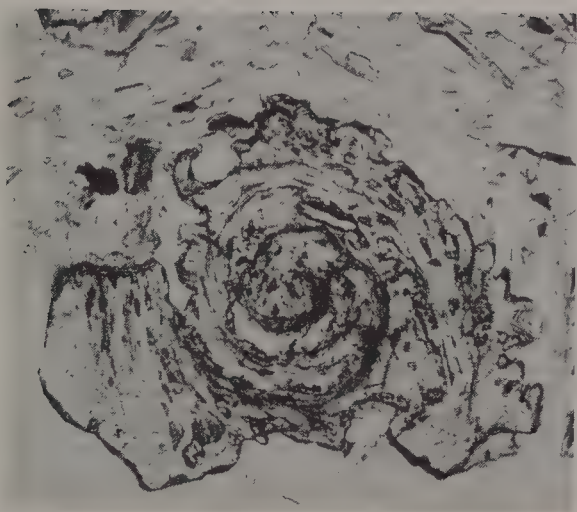


FIGURE 1. Snowball garnet—the original crystal figured by Flett (1912); PPL, $\times 55$.

enclosing matrix during metamorphic growth (Spry, 1963, 1969). The typical internal surface has an S- or Z-shape when viewed along the axis of rotation, but oblique sections show other shapes (Powell and Treagus, 1970).

The texture is syntectonic and must be distinguished from a somewhat similar appearing helicitic texture which is post-tectonic; curved helicitic textures are not uncommon in albite, chloritoid, staurolite, and andalusite.

The earliest explanation was that the crystal grew in a medium in which deformation was by simple shear (Spry, 1963) and growth took place by a process analogous to a snowball rolling down a snow-covered slope. However, this hypothesis appears incompatible with some evidence and details of origin are in doubt (Ramsay, 1962; Cox, 1969). The structure has also generally been called "rolled" or more specifically, rotational if the apparent rotation is less than 90° and snowball if it is greater. Powell and Treagus (1970) use the term rotational to cover all cases; this is preferable, as the amount of rotation is difficult to determine and the few snowball cases (such as the classic example of Flett (1912) in Fig. 1, interpreted as 540°) are questionable.

Analysis of the form of the spiral indicates the relationship between relative rates of growth and rotation. The amount of deformation of the enclosing rock can be estimated if certain assumptions are made (Spry, 1963; Wilson, 1972), but Cox (1969) and Powell and Treagus (1970) give reasons to doubt these. The importance of rotational garnets in the interpretation of structural and metamorphic evolution has been documented by many authors (e.g., MacQueen and Powell (1977) and references therein).

ALAN SPRY

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- Cross-references: *Gneiss*; *Metamorphic mineral growth*; *Nucleation of metamorphic minerals*; *Polymetamorphism*; *Polyphase metamorphic mineral growth*; *Porphyroblast*; *Schist*.

SODALITITE

Sodalitite is a coarse-grained igneous rock mainly composed of sodalite, with smaller proportions of acmite, eudyalite, and alkali feldspar. A *sodalithite* is an effusive rock with sodalite the only light-colored mineral and in which no olivine is present.

D. R. BOWES

Cross-references: *Alkaline rocks—undersaturated*; *Häuyinitite*; *Leucitite*; *Melilitite*; *Nephelinitite*.

SOLFATARA

A *solfatara* is a fumarole issuing volcanic gases with relatively high concentrations of sulfuric compounds, mainly H_2S . Since practically all fumaroles issue some sulfuric compounds, the solfataras do not form a very distinct group. The classical example is the Solfatara at Pozzuoli in the Phlegrean Fields near Naples, Italy. This solfataras is a volcanic caldera that last erupted in 1198. The term *soufrière* is also used particularly in the W Indies.

Solfataras are present in many high-temperature thermal areas, mainly in those associated with the volcanism of the great orogenic regions, e.g., in the circum-Pacific belt. Many dormant volcanoes of these regions have fumaroles of the solfataric type in the central craters or on their flanks. The postvolcanic activity is referred to as the solfataric stage, which may last for a considerable time after the last eruption. Well-known examples are Mount Lassen in N California, U.S.A., which was active in 1914–1919 and Popocatepetl in Mexico. Moreover, about 10 of the 40 active volcanoes in the Central American belt are now in the solfataric stage and the same applies to a great number of volcanoes in the other active belts of the world.

Postvolcanic solfataras are less common in the volcanic regions of the oceanic basins and flood basalt areas, but examples include the Askja caldera, Iceland (active in 1876 and 1961).

In general, solfataric activity appears mainly to be associated with volcanoes that have erupted silicic lava or tephra in great amounts. The main reason is that silicic lava carries more gases in solution than the mafic types of lava. The solfataras are the outlets for the gases expelled by the cooling lava.

The temperature of solfataras ranges from one hundred to several hundred °C, the highest values being observed in the cases associated with very recent volcanic activity. The predominant chemical component of the solfataric fumes is steam, in most cases making up more than 90 vol.%. The gases associated with the steam are mainly CO₂, H₂S, and H₂, with minor quantities of CO, SO₂, HCl, HF, and CH₄. CO₂ is often the predominant component, but cases are known where H₂ makes up about 50 vol.%, e.g., in the Namafjall area, Iceland. The chemistry of the gases is quite complex and their origin only poorly understood.

The H₂S is oxidized upon coming into contact with the air and elemental S is continuously being deposited around solfataras. In some cases the S deposits have been mined, e.g., in Italy, Iceland, and Japan.

G. BODVARSSON

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Cross-references: *Calderas and volcano-tectonic depressions; Explosive volcanic eruptions—classification; Fumarole; Geyser; Hot ground; Hot spring; Thermal activity; Volcanoes and volcanism.*

SOLID SOLUTION—See Vol. IVA

SÖVITE

Sövite is a coarse-grained calcite-rich carbonatite with the type locality at Lake Nordsjø in the Fen area, S Norway (Brögger, 1921); the rock is named after the Söve agricultural college. The corresponding fine-grained rock is termed *alvikite*. Brögger, who was the first to refer to a calcite-rich rock as being of magmatic derivation, gave special names to many mixed carbonate-silicate rocks, but Saether (1957) simplified them to pyroxene sövite and biotite sövite.

In its type locality, sövite often exhibits flow structure (Primäre Fluidalstruktur of Brögger) and contains 75–95% calcite. A typical composition is calcite (80%), apatite (7%), magnetite (3%), pyrite (1%), pyrochlore (0.5%) together with phlogopite, amphibole, aegirine, and accessory zoisite, fluorite, topaz, corundum, zircon, and rutile (Bjørlykke and Svinndal, 1960).

C. OFTEDAHL

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Cross-reference: *Carbonatite*.

SPILITE

The term *spilite* was introduced no later than 1819 by Alexandre Brongniart to denote aphanitic rocks containing nodules or amygdaloids of calcite or silica minerals. It has since undergone various changes of meaning (Vallance, 1960). Many authors of the present century have followed Dewey and Flett who in 1911 (see Vallance, 1960) described spilites (including amygdaloidal spilite with local names such as dunstone) as basic igneous rocks richer in Na and H₂O but poorer in K than basalts and, again unlike basalts, bearing albite as the typical feldspar with pyroxene or, more commonly, phases such as chlorite, epidote, calcite, or hematite. Subsequent studies, however, have cast doubt on the adequacy of any particular chemical diagnostic features. Individual spilitic bodies embrace contrasted, complementary compositions and in those varieties termed *K-spilite* (or *poeneite*) adularia rather than albite is characteristic. The definition adopted here broadly follows Cann (1969): "spilites are...rocks of basaltic texture in which greenschist facies mineralogy is completely or almost completely developed."

Occurrence and fabric characters

Spilites occur as bodies having the morphology of mafic lava flows or shallow intrusions. Most belong to subaqueous (especially submarine) associations, although subaerial examples are known. Fabric analogies with basalts exist at both meso- and microscopic scales (Amstutz, 1968) and, like quenched basalts, finer-grained spilites have notably variable textures. Unfilled joints and vesicles in pristine basalts have their counterparts in the veins and amygdaloids of spilites. Geometrically similar pillow structures occur in both basalts and spilites and both may show evidence of brecciation during eruption. But whereas in basaltic pillow lavas the selvages tend to be glassy and brecciated products include hyaloclastites, spilitic counterparts are finely crystalline and commonly bear signs of devitrification.

Spilitic rocks have worldwide distribution; examples range in age from Archaean to late Tertiary or Pleistocene but none has been reported among products of present-day volcanism. They are commonest in orogenic belts, where they may occur with clastic sediments rich in volcanic detritus or, in places, with radiolarites. Other spilites in orogenic zones are found with ultramafic and coarser mafic rocks (the so-called ophiolite association) or with keratophyres or quartz keratophyres (members of Dewey and Flett's "spilitic suite")—rocks

having fabric analogies with andesites or dacites and rhyolites. Less commonly, spilites have been noticed in epicontinental or continental associations. Ore deposits, notably of Fe, Mn, Cu, Pb, and Zn may occur with spilites. Well-known examples are those of the Olympic Peninsula and the Lake Superior regions, U.S.A.

Chemistry and mineralogy

Spilites are commonly heterogeneous in chemistry and mineralogy, with a tendency to almost monomineralic patchiness being most evident in brecciated spilites and in the selvages of spilitic pillows. The sorts of diversity found in such pillows are taken to be representative.

Unlike the glassy selvage of a basaltic pillow that approximates chemically to its more crystallized core, crystalline selvages of spilitic pillows differ markedly from their associated cores. Chlorite (mostly pycnochlorite to diabantine) is the commonest, in some cases almost the only, selvage phase. In some localities, however, selvages consist largely of such hydrous CaAl-silicates as epidote, pumpellyite, hydrogarnet, or prehnite, or of carbonates or hematite. By contrast, the cores of spilitic pillows carry albite (of transitional intermediate to low structural state) occupying textural sites that match those taken by labradorite in basalts. This albite, furthermore, exhibits twinning features (width of lamellae and frequency of repeats on the Albite law) more like labradorite than, say, vein albite. Typical cores consist of such assemblages as albite–chlorite, albite–chlorite–epidote, albite–chlorite–clinopyroxene (of normal basaltic types, although pigeonite is rare), or albite–chlorite–calcite. Albite being the dominant sodic phase in spilites, it follows that only albite-bearing parts can attain those chemical characters deemed diagnostic by Dewey and Flett.

As Amstutz (1968) has suggested, biased sampling may lie behind the seeming abundance of sodic spilites. Certainly, richness in Na is less apparent when the various parts of heterogeneous spilitic bodies are considered. Comprehensive sampling, indeed, suggests that the bulk composition of such a body, apart from enhanced content of H₂O and CO₂, may approach that of a basalt or a basaltic andesite (Smith, 1968).

Genesis

The characters of spilites clearly require a magmatic ancestry, but the particular nature of that ancestry and the status of spilitic mineral

phases have been much disputed. Dewey and Flett regarded these rocks as derived from a sodic (spilitic) magma with the observed mineral associations generated by late-magmatic, secondary action. Others have invoked spilitic magma but claimed a primary status for albite. The concept of spilitic magma, however, has failed to hold support; petrologists now favor a basaltic parentage. In this regard there have been two contrasted views, one igneous, the other metamorphic.

The igneous model holds that spilites are formed directly, by hydromagmatic crystallization of basaltic melt under high water-pressure (see Amstutz et al., 1974) by an extension of the magmatic trend that produces oligoclase basalts (mugearites). By such means, it is argued, a primary assemblage of albite-chlorite could become stabilized as end-members of the continuous and discontinuous reaction series. Pyroxene present, which may show curvaceous or branching habit, is interpreted as a metastable phase (Battey, 1956). There is a paucity of convincing experimental evidence to support this, and the apparently subophitic relations between demonstrably basaltic pyroxenes and relatively well-ordered albite in certain spilites is difficult to explain. For such textures to have formed, development of albite at the expense of some earlier phase is required. The distinctive character of spilitic albite and the occasional presence in it of calcic feldspar relics point to its being pseudomorphous after labradorite or a kindred plagioclase.

Clear evidence of such pseudomorphous replacement and of the secondary generation of spilite from solidified basalt comes from associations documented by Cann (1969) and Vallance (1974) in which parent, product, and intermediate stages are preserved. It follows that spilites inherit their fabric geometry from parent basalts and that spilitic mineral phases, apart from metastable pyroxenes and perhaps in some cases opaque oxides, have a secondary status resulting from adjustment of the solid parent to low-temperature, hydrous conditions. Such adjustment could arise by late-magmatic alteration, but the common evidence from fabrics of rapid quenching and brecciation serves to negate this as a general cause. Rather, it is probable that most spilites result from postmagmatic hydrothermal action, as occurs in burial or ocean-floor metamorphism.

Formation of low-grade metamorphic mineral assemblages in volcanic rocks by reaction with ambient pore fluid—termed degradation by Vallance (1969)—involves exchange of components between solid and fluid and metasomatic transfer. The final solid products are part

hydrolysate and part precipitate from cation-enriched solution. In the case of spilitization, chlorite appears to be the commonest hydrolysate phase and the fixing of displaced Na^+ ions from solution is determined largely by preference to exchange at primary feldspar sites. An analogous case of degradation is the development of propylite by wall-rock alteration in mineralized areas.

That spilitic degradation is a metamorphic process is endorsed by Turner (1981) and gains support from the experiments of Moody et al. (1983), who consider that spilites are merely metabasalts and that the older term is unnecessary. If that view were adopted, the distinctive pseudomorphous character of spilites would be lost among mafic hornfelses and schists. Such varieties of texturally-reconstructed metabasalts are recognized; so also should those in which the fabric geometry of a parent basalt persists through the operation of static degradational metamorphism.

T. G. VALLANCE

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- Cross-references: *Basalt*; *Burial metamorphism*; *Greenschist facies*; *Lava and lava eruptions*; *Ocean-floor metamorphism*; *Ophiolite*; *Reaction principle and reaction series*; *Volcanic rocks*.

SPINIFEX TEXTURE

The distinctive *spinifex texture* is formed by skeletal, platy, or acicular crystals of olivine,

orthopyroxene, or clinopyroxene, or their pseudomorphs in ultramafic and mafic lavas or silicate-rich furnace slag. In lavas, these crystals form a pattern on weathered surfaces which Australian geologists compared to the local desert grass *Triodia spinifex* (Fig. 1a). In Canada, the term is synonymous with "chicken track," "bird's foot," "bird track," "herring bone," "feather," or "lath" textures. The large lath-like or platy crystals, or skeletal, branching or acicular bundles of olivine or pyroxenes are set in a matrix of irresolvable material interpreted as devitrified glass (Fig. 1b). This groundmass may contain hopper-faced olivine phenocrysts, skeletal, cruciform magnetite and a variety of feathery crystallites. The complex skeletal or platy crystals may show random orientation, or may display a preferred arrangement. The texture may also exist on two scales, where large laths and needles of olivine or pyroxene enclose a groundmass that itself contains smaller, complex skeletal intergrowths of olivine or pyroxene (see Donaldson, 1982, for review and extensive bibliography).

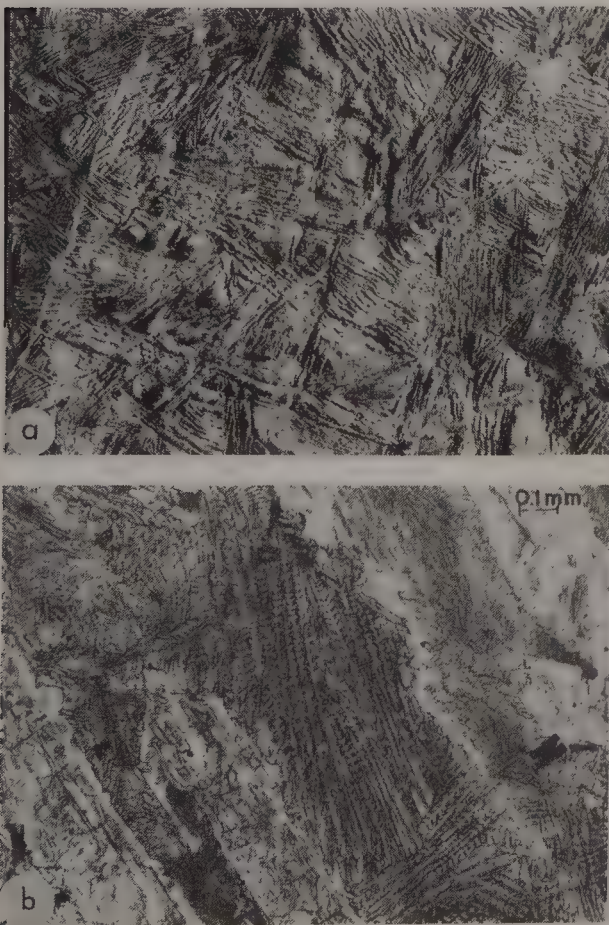


FIGURE 1. Spinifex texture in an ultramafic flow, Munro Township, Ontario, Canada: (a) field occurrence ($\times 0.6$); (b) photomicrograph showing bundles of clinopyroxene, some feather-like, and devitrified glass as infillings between large olivine blades. (From Pyke et al., 1973.)

In typical komatiite the spinifex-textured material is frequently altered or metamorphosed and the primary phenocrysts are replaced by pseudomorphs of amphibole (usually tremolite), talc, serpentine minerals, carbonate, or chlorite. In such metamorphosed ultramafic lavas, spinifex texture has been confused with ophicalcite or ophidolomite, where the acicular olivines, orthopyroxenes, and amphiboles showing ophitic-like texture were produced by recrystallization.

The true spinifex texture is the result of rapid cooling (but not quenching) of a high-temperature ultramafic lava. Similar textures are reported in intrusive rocks like the basal sills of the Bushveld Complex, and similarities to the cumulate textures of some mafic intrusions have been noted.

ADRIAN F. PARK

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Cross-references: *Archaean volcanism*; *Cumulate textures*; *Textures of igneous rocks*; *Ultramafic lavas*.

STATISTICS IN PETROLOGY

With the widespread use of electronic computers and calculators, the development of computer data banks, and the more abundantly obtained geochemical data supplied by various instrumental methods of analysis, elaborate processing and especially modeling of petrochemical data has led to the general application of standard statistical tests and treatments. Details of these procedures can be obtained from Koch and Link (1970–1971), Till (1974), Williams (1984), Davis (1986), and other texts. It is particularly important to have statistical statements of the reliability of experimental information used to construct the stability fields of igneous and metamorphic minerals and mineral assemblages, and the statistical error attached to radiometric age determinations.

The arithmetic *mean* of a set of numbers is the arithmetic average obtained by adding the numbers together and dividing by the number of separate values summed. This is the most

commonly used procedure, for example, in a series of optical measurements or in obtaining a chemical value from several repeated determinations. The *mode* is the most frequently occurring value and will often be near to the mean. By the calculation of *variance*, the spread of a distribution of data can be established and is calculated by adding the squares of the differences between each piece of data and the average of all the data and then dividing this sum of the squares by the number of items of data. If data show a single clustering around the mean or the mode, then, as a generalization, about two-thirds of the data lie less than one *standard deviation* (which equals the square root of the variance) away from the mean, while 95% of the data lie less than two standard deviations from the mean and less than 1% of the data lie more than three standard deviations from the mean. From standard deviation and the *coefficient of variation* (the ratio of the standard deviation to the mean) the *precision* (i.e., the reproducibility of a result) and sometimes the *accuracy* (how near to the truth the data are) can be assessed objectively. This enables true phase petrological and petrochemical deductions to be more readily distinguished from those arising solely from errors in the experiments or analyses. Thus, if a granite has $68.11\% \pm 0.30\%$ SiO_2 , where the standard deviation is ± 0.30 , then clearly it is unlikely that another granite with $68.32\% \pm 0.30\%$ SiO_2 is detectably different in silica by the technique used. On the other hand, a granite with $70.32\% \pm 0.30\%$ SiO_2 is significantly different. Unless some assessment of the likely errors in any measurement of an experimental phase diagram can be made, any deductions based on the diagram are hard to appraise objectively.

Correlation coefficients enable the degree of correlation between two elements in a rock or mineral series to be assessed precisely, whether it is positive or negative, whether correlation is perfect (+1 or -1), or whether there is no correlation (zero). *Discriminant analysis* enables overlapping multivariate data (i.e., data with more than two variables) to be sorted into a number of classes. For example, the distinction between a gneiss derived from an igneous rock and a gneiss formed from a sediment may be possible if enough characterization of each type can be previously established; samples of unknown origin may then be more confidently assigned to one of the two possibilities. *Trend surface* maps of petrochemical parameters, in which the computer fits equations of varying complexity from linear to perhaps even sixth-order equations to the data, have become possible. These have been widely used in

geochemical prospecting, where the search for anomalously high or low concentrations of chemical elements and the associations of these anomalies are important as indicators of ore deposits. Trace-element excesses or deficiencies in rocks, soils, and plants also have important biomedical implications. *Factor analysis* enables the significant variables to be unraveled from complex multivariate data of the kind that chemical analyses of silicates inevitably are. A *t-test* enables an estimate of the probability of a set of data belonging to a given group or not, taking into account the mean values of the sample and of the group, the standard deviation attached to the group mean, and the number of determinations upon which the group mean is based. *Cluster analysis* (Howarth and Leake, 1980) identifies groupings or specific subsets in a large data matrix.

A great many petrologic manipulations involve plotting one variable against another on a simple graph and drawing or calculating a line that is the best fit through the points. How "the best fit" is defined if the points lie perfectly or nearly perfectly on a straight line is not important, but very different results can be obtained if the data are widely scattered, depending upon the method used to obtain the best-fitting line or *regression* line. If x and y are plotted against each other (Fig. 1), a line that gives the minimum possible total deviation of all the points from the line measured along the x direction will be different from the line that gives the minimum possible total deviation along the y direction, even though the same data points are used. The *reduced major axis* method of fitting a line to points is much more satisfactory, as this method calculates a line that

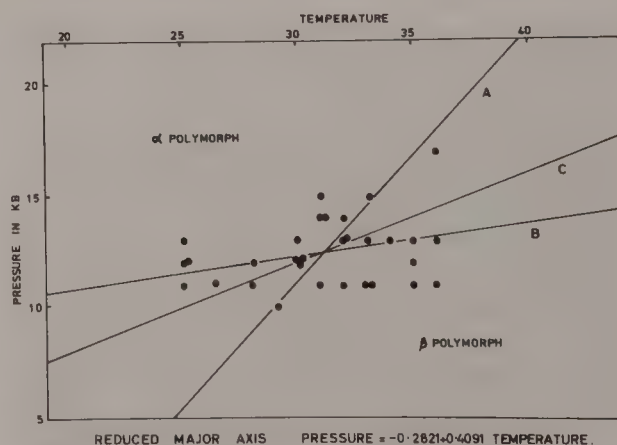


FIGURE 1. Three different relationships deduced from the same data depending upon the method of fitting the line to the plotted points. Line A is a regression of temperature on pressure; line B is a regression of pressure on temperature; and line C is the reduced major axis method.

minimizes the total deviation of both x and y considered together.

Over the past ten years there have been major advances in the use of computing techniques for modeling of major-element, trace-element (including rare-earth element), and isotopic data (e.g., Miesch, 1976; Stormer and Nicholls, 1978), particularly to constrain source areas for magma generation and to distinguish groups of rock types (see **Geochemical modeling in petrogenesis**).

BERNARD E. LEAKE

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Cross-references: *Geochemical modeling in petrogenesis*; *Petrochemical calculations*; *Trend surface analysis in petrology*.

STEINMANN TRINITY

The so-called green rocks (“Grüngesteine,” “roches vertes,” “pietre verdi”) have attracted the attention of geologists ever since the days of H. B. de Saussure, L. von Buch, and B. Studer in the late 18th and early 19th centuries (see Dietrich, 1969). Regular green rock associations were composed of two or more of the following ultramafic and mafic rock types: peridotite, serpentinite, gabbro, diorite, amphibolite, dolerite (diabase–porphyrite), basalt, basaltic tuff, prasinite, spilite and variolite. Their abundance in the Alps, and in other recumbently folded and imbricated orogenic belts, was recognized at an early stage, along with their occurrence restricted to specific stratigraphic levels and tectonic settings. The term ophiolite, initially the Greek-derived synonym for serpentinite, was later expanded to cover just such green rock associations of the alpine type.

Gustav Steinmann, early this century, was among the first geologists to grasp the essentials of the ophiolite suite and the significance of its palaeogeographic and geotectonic context. Studying the green rocks and their associated sediments in the Iberg Klippen east of Lake Lucerne, Switzerland, in the Rhaetic (now Platta) nappe of the Grisons and the Engadine, in the Tuscan and Ligurian Apennines of Italy, and later in many other places, he arrived at several important conclusions.

1. Three members of the green rock family, viz., serpentinite or peridotite, gabbro and spilite–variolite (often pillow lavas)–dolerite or diabase–porphyrite dykes, form ophiolite suites that occur in specific zones of the younger fold belts. Their relationships are shown, in cartoon form, in Fig. 1.
2. Ophiolite suites are restricted to specific fold- or thrust-nappes, such as the Upper Penninic and some of the Austro-alpine nappes of the Swiss Alps, and the Liguride nappe of the Apennines. They do not transgress their associated sediments and must therefore be regarded as penecontemporaneous and of the same palaeogeographic facies. Their tectonic behavior was passive and they were not, as E. Suess claimed, actively injected along the soles of nappes and thrust-slices.

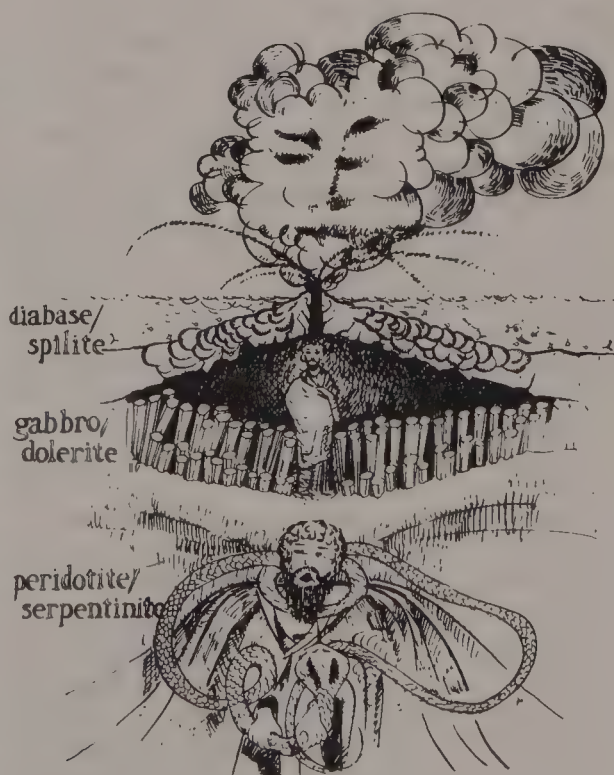


FIGURE 1. The Steinmann Ophiolite Trinity. (From Coleman (1977) after unpublished drawing by E. Den Tex.)

3. Ophiolite suites nearly always occur in close association with rather thin accumulations of three types of sediments interpreted as having been formed in a deep-sea environment: (a) coarsely banded, whitish, light-gray and reddish limestones, usually overlying the radiolarites and containing tiny pelagic foraminifera such as the tintinnid *Calpionella*, as well as ammonite aptychi or rhyncholiths; (b) radiolarites and variegated abyssal clays; and (c) clays and weakly argillaceous limestones, interstratified in clays in the form of banks and nodules, underlying both (a) and (b).

Steinmann also inferred that ophiolites are intrusive igneous rocks emplaced as very large plate-like bodies, or placoliths, in the suture between basement and sedimentary cover in the border zone of an alpine-type geosyncline. Thus, unwittingly, he laid the foundation for the modern concept of an oceanic lithosphere plate with its characteristic mid-oceanic ridge association caught up in an alpine-type orogenic belt, or "Petasmorogen" as he chose to call it. However, Steinmann denied the now obviously volcanic character recognized by some of his Italian and British contemporaries, of certain ophiolite constituents like pillow lavas, variolites, hyaloclastites, breccias, tuffs, and ashes (references in Grunau, 1965). He also failed to appreciate the cogenetic relation between ophiolites and associated sodic granites, although he admitted such a relationship with Cu-bearing quartz veins (Steinmann, 1927, and earlier papers referred to therein).

It was E. B. Bailey who coined the term "Steinmann Trinity." His work on the Ballantrae complex in SW Scotland and on the ophiolites of Anatolia in Turkey had convinced him of the soundness of most of Steinmann's ideas, especially the association of ophiolites with radiolarites and related rocks. But the profane trinity he dedicated to Steinmann was of a somewhat hybrid nature. It consisted of two ophiolite members (*serpentinite* and *pillow lava*) and one sedimentary member (*radiolarian chert*).

Bailey strongly supported the case for effusive and explosive volcanic activity, and for alteration of their products in contact with sea water, even to the extent of interpreting most of the serpentinites as ultramafic lavas. He also introduced the term Ankara Mélange (to be superseded by Gansser's Coloured Mélange) for large-scale, allegedly tectonic, breccias containing fragments of every conceivable rock type of the ophiolitic eugeosynclinal association. It is now known that they are essentially sedimentary or tectonic megabreccias, or olistostromes,

that herald the onset of tectonization and emphasize the development of a steep continental slope to delimit a deep-sea trench (Bailey and McCallien, 1960, and earlier papers referred to therein).

Although the Steinmann Trinity of Bailey became firmly established in the Anglo-Saxon geological literature, it never obtained a proper footing on the European Continent. H. M. E. Schürmann and Kündig were among its few advocates, while the principal critics cast doubt on the abyssal character of the radiolarites and related sediments, and attempted to reinstate Steinmann's original ophiolitic triad, viz., *serpentinite* or *peridotite*, *gabbro*, and *dolerite* (*diabase*) or *spilite*, or to add further refinements to its sedimentary context.

Vuagnat (1963) introduced the French term "*trilogie*" for the original green rock triad, but emphasized the volcanic character of the dolerite-pillow lava-spilite-variolite-breccia member as distinct from the plutonic character of the other two members.

Grunau (1965) suggested splitting Bailey's hybrid trinity into a Steinmann Ophiolite Trinity, as defined above, and a Steinmann Sedimentary Trinity consisting of radiolarian cherts and abyssal clays, tintinnid-bearing limestones, and shales with interbedded argillaceous limestones. He also introduced a rather complex systematic nomenclature for ophiolites and their associated sediments. Carefully weighing the available evidence, Grunau concluded that Kündig's separation of the sediments into a chert-pelagic limestone-siliceous shale group, characteristic of the abyssal plain, and a catastrophic olistostrome-turbidite (*mélange*-*flysch*-*graywacke*) group, characteristic of the slope-trough hinge, is essentially correct, and that their association with ophiolites is worldwide and recurs periodically in the early stages of orogenic cycles.

Recent developments in knowledge of the oceanic lithosphere, its creation, its spreading in the form of global plates, and its subduction at seismically active continental margins, has brought the trinity back into the geologic limelight. Rock associations corresponding closely to both the ophiolitic and the sedimentary Steinmann trinites have been detected in the ocean floor at ridges as well as in deep-sea trenches. Trinity associations thus appear to represent sections of oceanic lithosphere incorporated in orogenic belts by under- and overthrusting. The regular superposition, sometimes repeated, of inorganic abyssal clays on organic oozes, chert and chalk may be due to their deposition either below or above a carbonate compensation depth changing with latitude, or else to periodic changes in the

submarine volcanic supply of carbon dioxide and silica (Grunau, 1965; Heezen and MacGregor, 1973; see also Coleman, 1977).

E. DEN TEX

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- Cross-references: *Alpine mafic magma stem*; *Basalt*; *Dolerite*; *Lava and lava eruptions*; *Mid-ocean ridge and ocean-floor petrology*; *Ocean-floor metamorphism*; *Oceanic crust*; *Ophiolite*; *Peridotite*; *Serpentinite*; *Spillite*.

SYENITE

Syenites are holocrystalline coarse- to medium-grained intermediate igneous rocks in which the dominant mineral is alkali feldspar. Most syenites are hypersolvus rocks in which the feldspar is micro- or cryptoperthitic alkali feldspar (“alkali feldspar syenite” of Streckeisen, 1976) but two-feldspar (i.e., subsolvus) syenites with albite–oligoclase and a discrete perthitic alkali feldspar phase also occur (syenite of Streckeisen, 1976), grading into monzonite. Quartz is often present and if $Q > 5\%$ the rock is called quartz syenite (sometimes *nordmarkite*) which grades in turn ($Q > 20\%$) into granite. Slightly undersaturated (nepheline-bearing) syenites are also a common type. Essentially monomineralic alkali feldspar syenites are called *perthosites*. With increasing anorthite content, hypersolvus syenites grade into *syenodiorites* (monzonite); intermediate types on this trend containing ternary alkali feldspars are often called *larvikitic syenites*. Such rocks often contain the strikingly iridescent alkali feldspars well known from the type

larvikite, which is a syenite with minor nepheline. Syenites are often distinguished on the basis of their mafic constituents, of which augite–fayalite syenite is an important example. *Shonkinites* are mafic syenites with pyroxene $>$ alkali feldspar. Syenites may be divided (essentially on the basis of mafic constituents) into calc-alkaline and alkaline types, of which the latter predominante (sometimes distinguish as *soda-syenites*). The main extrusive equivalent of syenite is trachyte. The name syenite is an ancient one, in use at the time of Pliny, although the rock from Syene, Egypt, is, in fact, a hornblende granite (see Johannsen, 1937).

Mineralogy

Feldspars. Some complexity is introduced into early schemes of classification (such as Johannsen, 1937) of alkali-feldspar-rich rocks (syenites, granites, and their extrusive equivalents) by lack of understanding of the nature of the feldspar variation exhibited in such rocks and absence of a satisfactory nomenclature for alkali feldspars. Recent understanding of these relations stems largely from the work of Tuttle and Bowen (1958), who comment on this early nomenclatural complexity. In the Streckeisen (1976) classification (based on *modal* feldspar) this difficulty persists in simplified form, because the majority of syenites possess restricted normative feldspar mineralogy (see Morse, 1969) and yet, depending on the actual nature of the feldspar phases, might be included under monzonite, syenite, or alkali-feldspar syenite. Most syenites are hypersolvus rocks containing a single set of discrete feldspar crystals that are themselves crypto- or microperthitic intergrowths of K- and Na-rich phases. These may be microcline–low-albite, orthoclase–low-albite, or, particularly in feldspars with significant anorthite component (syenites of “larvikitic” type), sanidine–anorthoclase cryptoperthites. It is essential in descriptions of feldspars in syenitic and allied rocks to give an estimate of the *bulk* feldspar composition as well as that of the unmixed phases.

Plagioclase as a distinct crystal species (i.e., not in perthitic intergrowth) may appear in syenites for a variety of reasons. Two-feldspar (i.e., subsolvus) syenites may develop by direct magmatic crystallization at high water-vapor pressures; this is possible at lower water-vapor pressures if appreciable anorthite is present. Calcic syenites of this type would range into *monzonites* (or *syenodiorites* if showing alkaline tendencies). Plagioclase (andesine) may also appear as cores to perthitic feldspar crystals in rocks that initially crystallized plagioclase but that terminated crystallization with a hypersol-

vus single-phase alkali feldspar. This is most common in calcic syenites, but rare grains with plagioclase cores are almost always encountered in hypersolvus rocks. A late, interstitial albite phase is common in many hypersolvus syenites, and may be produced by unmixing or by late sodic fluids. Finally, a two-feldspar syenite might evolve from a single-feldspar assemblage by recrystallization contingent upon the movement of a crystal mush or tectonic deformation.

Other minerals. Minor interstitial nepheline or quartz may occur, often in rather variable proportions, and in some cases leading to a smooth gradation into quartz syenite or nepheline syenite. Calc-alkaline syenites may contain biotite or hornblende like the granites to which they are otherwise similar. Alkaline syenites contain a wide variety of mafic phases. Pyroxenes, usually Mg-poor, may range from diopsidic augites to ferrohedenbergite, and in peralkaline syenites to acmite. A late acmitic pyroxene is often overgrown onto earlier augite or amphibole. Trends of pyroxene variation in a variety of syenites are summarized by Stephenson (1972). Amphiboles range from hastingsite to riebeckite-arfvedsonite. Fayalite is commonly present, often altered, together with magnetite-ilmenite intergrowths; Fe-rich biotite is sometimes present, as fringes on ores, or in calcic syenites as an essential poikilitic phase. Apatite, zircon, fluorite, allanite, and a wide range of rare minerals are accessory minerals in syenitic rocks.

Distribution and genesis

Calc-alkaline syenites sometimes occur as marginal variants of granodioritic plutons. Alkaline syenites (and their extrusive equivalent trachyte) are the most abundant rock type in many continental alkaline provinces (see Bailey and Schairer, 1966). Characteristically they form central plutonic complexes of relatively small size (up to 35 km maximum, in the Gardar province, S Greenland—Emeleus and Upton (1976), but characteristically much less). Saturated to slightly oversaturated syenites are commonly intimately associated with alkali granites, or undersaturated syenites with nepheline syenites; Bailey and Schairer (1966) discuss mechanisms by which these associations may evolve. Alkali gabbroic magma is also commonly associated with syenites, although the relative volumes of basic and syenitic rocks vary immensely from one province to another. The most important structural setting is one of crustal extension, such as the present African rift systems or the ancient Gardar rift. Syenites are generally derived by fractionation of alkali basaltic magmas generated in regions of the

upper mantle enriched in alkalis, certain trace elements, and volatile constituents (see **Alkaline rocks—undersaturated**).

Syenite genesis is reviewed in Fitton and Upton (1987).

Other terms

Akerite—an augite-bearing syenite (Aker, Norway).

Bigwoodite—an alkali syenite with orthoclase, microcline and albite.

Birkremite (*bjerkreimite*)—a light-colored quartz-alkali feldspar syenite with some hypersthene (included in the charnockite suite by some authors); a hypersthene-bearing kalialaskite (Birkrem, Norway).

Calcisyenite—a syenite containing labradorite-bytownite rather than andesine-oligoclase as its accessory plagioclase.

Cancarixite—an aegirine-augite-bearing alkali quartz syenite with lamprophyric affinities.

Deldoradoite—a leucocratic cancrinite syenite (Deldorado Bay, Colorado, U.S.A.).

Drakonite—a biotite syenite containing oligoclase to labradorite and sometimes minor sodalite and aegirine-augite (Drachenfels, W Germany).

Elkhornite—an alkali feldspar syenite with some labradorite (Elkhorn Mtns., Idaho, U.S.A.).

Evergreenite—a nordmarkite containing sulfide ores (Evergreen Mine, Colorado, U.S.A.).

Finandranite—a K-rich syenite with microcline, amphibole, biotite, ilmenite, and apatite.

Hatherlite—an orthoclase-bearing biotite-hornblende syenite.

Hedrumite—a leucocratic alkali syenite.

Hortite—a dark-colored syenite possibly formed from gabbro by limestone assimilation.

Lakarpite—a syenite composed of orthoclase or microcline, calcic plagioclase, and an arfvedsonite-like amphibole.

Leeuwfonteinite—a synonym of hatherlite (Leeuwfontein, Transvaal, S Africa).

Marosite—an intrusive rock composed of biotite and augite with subordinate feldspar and feldspathoid; compositionally between shonkinite and nepheline diorite (Pic de Maros, Indonesia).

Ordosite—a dark-colored acmite syenite.

Orthosyenite—a silica-saturated syenite.

Perthosite—an alkali syenite with minor aegirine-augite.

Plauenite—a potassic, plagioclase-rich syenite; often a syenodiorite (Plauen, E Germany).

Semejtaavite—a quartz-alkali feldspar syenite with anorthoclase, ilmenite, augite, and riebeckite.

Sölvsbergite—a micro-alkali syenite richer in mafic minerals than bostonite.

Sviatonossite—an alkali syenite containing orthoclase, perthite and albite, and andradite (Sviatoy Noss, Transbaikalia, U.S.S.R.).

Syenitite—a biotite syenite.

Syenoide—an abbreviation for foidal syenite.

Thuresite—a sodic amphibole-bearing alkali syenite.

Tönsbergite—an orthoclase- and andesine-bearing rock resembling larvikite (Tönsberg, Norway).

Yogoite—a syenite containing approximately equal proportions of orthoclase, plagioclase, and augite (Yogo Peak, Montana, U.S.A.).

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Cross-references: *Alkaline rocks—undersaturated; Igneous rocks—classification and nomenclature; Intermediate igneous rocks; Plutonic rocks—classification and nomenclature; Rift valley volcanism; Trachyte; Volcanic rocks.*

T

TEKTITE

Tektites are pieces of black or green translucent glass having a superficial resemblance to obsidian; they were first described by Meyer (1788). Their shape is of either a droplet or a spherule. Some have a more complex shape consistent with aerodynamic ablation or sculpting whilst molten. Their size varies from ca. 10 cm to <1 cm diameter. The glass differs chemically from obsidian in being virtually free of volatiles (even when frothy), free of microlites, and in containing lechatelierite (pure SiO₂ glass).

Although tektites do not contain microlites (crystals that grew in the glass) they do contain crystal fragments and other inclusions. These include Ni-Fe metal-sulfide spherules, globules of lechatelierite, baddeleyite, and zircon, and, less frequently, relict silicate grains. The glass matrix itself is usually inhomogeneous in composition and displays convolute layering. These features indicate a rapidly generated, quenched glass that did not remain molten long enough to homogenize or equilibrate. The melt appears to have been generated by impact, and quenched during passage through the atmosphere. Microtektites have all the above characteristics, but are much smaller—generally <1 mm in diameter.

Tektites occur in superficial deposits in well-defined geographical areas, and local names are given to them (*australites*—southern Australia; *bediasites*—Bidai (Bedias) Indians of Texas, U.S.A.; *billitonites*—Belitung (Billiton) Is., near Sumatra; *indochinites*—SE Asia; *moldavites*—R. Moldau (Vltava), Czechoslovakia). Microtektites have been found with tektites on land, but have been found in abundance in cores of deep marine sediments, often associated with coastal tektite-strewn fields.

K-Ar and fission track dating of both microtektites and tektites has permitted strewn fields to be defined; chemistry supports this. Three, possibly four such strewn fields have tentatively been identified.

The origin of tektites has proved problematical, and models are still controversial. That they are droplets of melt splashed out from meteorite impact sites is now generally agreed, but the

location of these sites is contentious: both the Earth and the Moon have been suggested. Compositions of tektites indicate gross similarities with average continental rocks, and particularly the soils derived therefrom, whilst contrasting sharply with known lunar rocks and regolith. Likewise, the lack of cosmic ray damage on the surface of tektites and microtektites does not suggest a prolonged residence in space or Earth orbit. The balance of evidence suggests a terrestrial impact origin, although the correlation of particular strewn fields with particular astroblemes, namely the Czechoslovakian field with the Ries Kessel crater in southern W Germany, and the Ivory Coast field with Bosumtwi crater, in Ghana, remains contentious.

ADRIAN F. PARK

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Cross-references: *Astrobleme*; *Lunar petrology*; *Meteorites—petrology*; *Shock metamorphism*; *Volcanic glass*.

TEPHRA

The term *tephra* (from the Greek for ash) was first used by Aristotle in an account of an eruption of Hiëra (Vulcano) in his book "Meteorologica." It refers to the clastic volcanic material that is transported from the vent through the air during an eruption (Thorarinson, 1954) and thus is a parallel term to lava, which signifies all the molten material flowing from a vent. It is particularly useful for general description of predominantly unconsolidated pyroclastic deposits (Schmid, 1981), as the term implies no particular grain size. Ash, on the other hand, has been specifically defined by Fisher (1961) as pyroclastic debris of diameter

<2 mm and pumice is restricted by many authors to pyroclastic debris that will float on water.

The composition of tephra may range from acidic to basic, depending on the host magma, and the content can vary from entirely primary fragments to entirely accessory fragments (Wentworth and Williams, 1932). Whether the term should be restricted to airfall pyroclastic deposits or used for both airfall and pyroclastic flow deposits (the products of nuées ardentes) is a matter of debate. However, in view of the reference by Thorarinsson (1969) to tephra flows it is considered that pyroclastic flow deposits can be called tephra if unconsolidated. Consolidated pyroclastic flow deposits can be called *ignimbrite* if they show welding of fragments or *sillars* if the fragments are unwelded. For detailed descriptions, subdivision is made according to the size of the major constituent grains, using the terms of Wentworth and Williams (1932) and the size divisions of Fisher (1961) (Fig. 1).

If tephra becomes indurated by compaction or cementation, or becomes weathered, it becomes *tuff* if the grain size is <64 mm. Where the average size exceeds 64 mm the rock is referred to as an *agglomerate* if rounded pyroclasts predominate or as a *pyroclastic breccia* if angular pyroclasts predominate (Schmid, 1981).

Because of the widespread distribution of most tephra layers, they are of great value for stratigraphic correlation, particularly during Quaternary and Recent times. An eruption marks a point of time in geologic history and the resultant tephra deposit, where preserved, provides a stratigraphic marker corresponding to this event. Where many eruptions have occurred during a period of time, the tephra sequence records a complete volcanic history of an area. For example, in the Rotorua-Taupo



FIGURE 2. Tephra section near Taupo, New Zealand, showing some of the tephra layers that have been erupted in the last 20,000 years.

district of the North Island, New Zealand (Fig. 2) more than 30 pyroclastic eruptions can be recognized in the last 20,000 years (Healy et al., 1964). The dates of many eruptions can be obtained by ¹⁴C dating of carbonaceous material associated with the tephra and the chronology established from the tephra units is called tephrochronology (Thorarinsson, 1944).

Tephra units are studied as single or multiple layers within a tephra column and correlated over as wide an area as possible to produce an isopach map. From these isopachs the source volcano of the tephra can usually be established; e.g., the Kaharoa tephra on the North Island of New Zealand, that was erupted approximately 600 years ago, can be shown to have a source vent on Mt. Tarawera (Fig. 3). For ultraplinian eruptions, however, isopach maps should be used with caution: the Taupo (New Zealand) pumice deposit is thickest 20 km downwind from the vent (Walker, 1980). Wind directions during an eruption can also be determined from the distribution of the tephra and commonly it is found that, because of a prevailing wind, a tephra layer may be concentrated on one side of a volcano. Changes of wind direction during the course of an eruption can also be demonstrated where there are two or more lobes of tephra (Fig. 3).

In Iceland, recent tephra deposits are included in loessic soils and peat horizons. As these accumulate rapidly, the eruptive events can be clearly distinguished (Fig. 4). Because of the frequency of eruptions (Hekla has erupted 15 times since 1104 A.D.—Thorarinsson and Sigvaldason, 1972) tephrochronology has proved invaluable in Iceland for studies of

GRAIN SIZE (mm)		TEPHRA		INDURATED TEPHRA		
—256—	COARSE	BLOCKS AND BOMBS	PYROCLASTIC BRECCIA AGGLOMERATE			
	FINE					
—64—						
LAPILLI			LAPILLI TUFF			
—2—						
—0.063—	COARSE	ASH	COARSE	TUFF		
	FINE		FINE			

FIGURE 1. Divisions of tephra based on grain size. (Modified after Fisher, 1961.)

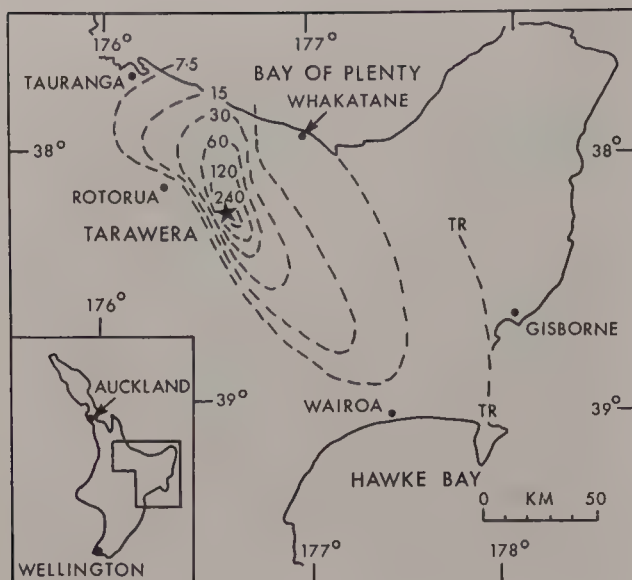


FIGURE 3. Isopach map showing the distribution of the Kaharoa tephra in the North Island, New Zealand; isopachs shown in centimeters; tr = trace. (After Healy et al., 1964.)

shoreline development, climate and vegetation history, glacier oscillations, and wind and water erosion, as well as for dating archaeological sites.

The lithological appearance of a tephra layer is not usually constant around the source volcano, or with distance from it, and thus correlation must frequently be made using diagnostic characters. Sometimes, a tephra deposit has a distinctive mineralogy (Cole, 1970), but more often the physical properties or chemical composition of the glass or minerals must be used. Trace-element ratios may also be used (Kohn, 1970), while for older tephtras fission-track dating (e.g., Seward, 1976) and

magnetostratigraphy (e.g., Verosub, 1981) have been used, in conjunction with mineralogy and chemistry, as bases for correlation.

Reviews of processes which form tephra and characteristics of tephra erupted and deposited in various geologic settings are provided by Self and Sparks (1981), Fisher and Schmincke (1984), and Cas and Wright (1987).

J. W. COLE

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Cross-references: *Explosive volcanic eruptions—classification; Lava and lava eruptions; Nuée*

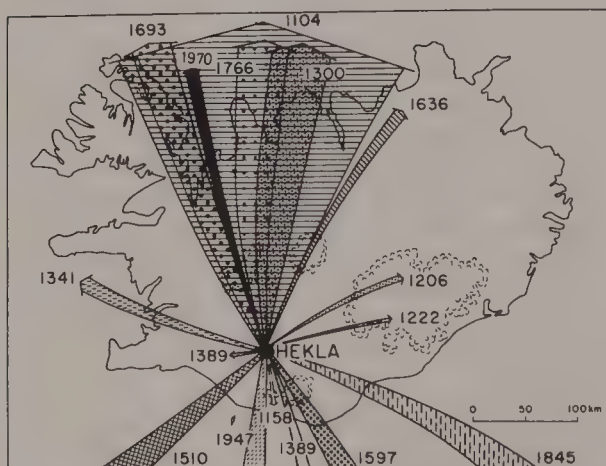


FIGURE 4. Map showing the direction in which the tephra was dispersed during the initial phase of each of the 15 eruptions of Hekla, Iceland, since 1104 A.D. The width of each arrow indicates approximately the relative estimated volume of the tephra layers. (From Thorarinsson and Sigvaldason, 1972.)

ardente; Palaeovolcanology; Pyroclastic eruptions; Pyroclastic flow; Volcanic glass; Volcanoes and volcanism.

TEPHRITE, BASANITE

A *tephrite* is an igneous rock of basaltic character that generally occurs as lava flows, but occasionally as minor intrusions. Its essential constituents are calcic plagioclase and clinopyroxene in association with nepheline, leucite, or analcime. With the addition of olivine to this mineral assemblage, the rock is termed as *basanite*, a term used by the Greeks of classical times to describe touchstones used in testing the purity of alloys of gold and silver. Both tephrites and basanites vary in color from gray to black. They are characteristically porphyritic with a very fine-grained groundmass. Some are holocrystalline but in many the groundmass is hypocrySTALLINE. Because of this the feldspathoids are commonly occult in the groundmass. Such rocks are often referred to as tephritoids and basanitoids with the norm calculated from a chemical analysis needed before they can be classified. Discrimination from alkali basalts, which may contain accessory analcime or nepheline, is generally made on the basis of >10% normative nepheline.

The calcic plagioclase varies from bytownite to labradorite; when present as phenocrysts it is commonly zoned. Titanaugite is the usual ferromagnesian constituent, being present as both phenocrysts and in the groundmass. Olivine, which normally is present only as phenocrysts, is an essential constituent of basanites. Alkali amphibole and biotite occur in some rocks. Apatite, opaque ores and perovskite are present as accessory minerals. With progressive reduction in the proportion of feldspar, the rocks grade into leucitite and nephelinite (Table 1).

Plutonic and hypabyssal equivalents of leucite tephrite are Essexite and microessexite and of nepheline tephrite are theralite and microtheralite. Mineralogically the hypabyssal types resemble the respective plutonic rocks.

The lavas of Vesuvius are the best known leucite tephrites. They are rich in leucite that

commonly occurs as euhedral phenocrysts. Basanites are a notable constituent of the Cenozoic volcanic assemblage of the Western Rift of Central Africa.

The following are terms used in relation to tephrites.

- Amphigenite*—a leucite tephrite.
 - Buchonite*—a dark-colored orthoclase-bearing tephrite.
 - Campanite*—a variety of leucite tephrite; also a pseudoleucite-bearing nepheline syenite.
 - Guardiaite*—a nepheline leucobasanite (Guardia, Italy).
 - Leucotephrite (leucitephrite)*—a tephrite with leucite as the only sodic feldspathoid.
 - Ordanchite*—an olivine-bearing haüyne-andesine-augite-hornblende tephrite (Banne d'Ordanche, France).
 - Ottajanite*—a leucite tephrite with phenocrysts of calcic plagioclase, leucite and augite (Ottajano, Mt. Somma, Italy).
 - Vesuvite*—a tephrite rich in leucite (Vesuvius, Italy).
- The following are terms used in relation to basanites.
- Caltonite*—a dark-colored analcime-bearing basanite with microphenocrysts of olivine and augite in a trachytic groundmass.
 - Heptorite*—a dark-colored haüyne basanite with a glassy groundmass.
 - Mandchurite*—a nepheline basanite.

D. R. BOWES

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Cross-references: *Alkaline rocks—undersaturated; Basalt; Leucitite; Mediterranean suite; Nephelinite; Rift valley volcanism; Volcanic rocks.*

TABLE 1. Main types of tephrites and basanites

	Without olivine	With olivine
With feldspar	Leucite tephrite Nepheline tephrite Analcime tephrite	Leucite basanite Nepheline basanite Analcime basanite
Without feldspar	Leucitite Nephelinite	Olivine leucitite Olivine nephelinite

TEXTURES OF IGNEOUS ROCKS

Description of igneous rocks requires information about crystallinity, grain size, shape of grains, and mutual relations of grains, in addition to that dealing with the composition of the constituent materials. The rocks can be entirely of crystals (*holocrystalline*), crystals and glass (*hypocrystalline* or *merocrystalline*), or entirely of glass (*holohyaline*). There is no unanimity concerning sizes for *fine*-, *medium*-, and *coarse-grained* rocks but <0.05 mm, 0.05–1 mm (or 2 mm—the IUGS recommendation) and >1 mm (or 2 mm) correspond generally with grain sizes of many effusive (volcanic), hypabyssal, and plutonic rocks, respectively (0.05 mm is the approximate size at which the human eye ceases to resolve individual particles). However, the variation in range of sizes used is illustrated by the ranges of <1 mm, 1–5 mm and >5 mm used by some authors.

Coarse-grained igneous rocks are referred to as being *phaneritic* (dycrystalline); the term for rocks whose crystal grains cannot be distinguished by the naked eye is *aphanitic* (eucrystalline). Where the crystals are too small to be distinguished under the petrographic microscope, the term *cryptocrystalline* (adiagnostic) is used.

Rocks may be *equigranular* or *inequigranular*, with those that exhibit the presence of relatively large crystals (*phenocrysts*, *megacrysts*) in a finer-grained *groundmass* (matrix, mesostasis) referred to as having porphyritic texture.

Grain shapes are referred to as *euhedral* (idiomorphic, automorphic) for those grains that exhibit crystal faces shown by geometric, regular outlines, particularly seen under the microscope. For those grains that exhibit only partial geometric outlines the term *subhedral* (hypidiomorphic) is used; for those grains that lack a geometric outline, the term *anhedral* (allotriomorphic, xenomorphic) is used.

Texture is dependent on the mutual relationships of the grains with the amount and shape of contact by the boundary outline, particularly as seen in thin section, being of major significance. Textural terms have proliferated, as shown by the following list, which excludes textures of cumulates (see **Cumulate textures**). This list also gives related fabric terms. More common textures are illustrated by photomicrographs 1–48 (1, 2 etc.) for which magnification is $\times 80$ unless stated otherwise; terms without a preceding number are not illustrated; PPL is plane polarized light, XN is crossed nicols.

1. *Allotriomorphic-granular* (*xenomorphic*—
2. granular), *saccharoidal*, *aplitic*—even-grained anhedral crystals; aplitite with

orthoclase, quartz and altered plagioclase; Weinduth, Massachusetts, U.S.A., XN (1); pyroxenite, Webster, N Carolina, U.S.A., XN (2).

3. *Amygdaloidal*—containing many amygdals; basalt with chlorite filling amygdals, Keweenaw Co., Michigan, U.S.A., PPL.

Aphanitic—crystalline components not distinguishable by the unaided eye (aphanite); both microcrystalline and cryptocrystalline textures are included—see *Felsitic*.

Aphyric—fine grained or aphanitic and without phenocrysts.

Arabesquitic—an apparently homogeneous groundmass in certain porphyritic rocks shows, under crossed nicols, irregular patches resembling arabesques.

4. *Axiolithic*—elongate radially disposed aggregates of quartz and feldspar arranged along a central axis; devitrified volcanic glass, Marysville, Utah, U.S.A., XN.

Basiophitic—a variant of ophitic in which the mesostasis is composed of augite; the mesostasis is referred to as *basimesostasis*.

Belonite—elongate or acicular crystallite with arcuate or pointed ends.

Bimagmatic—two generations of minerals occurring in a porphyritic rock.

5. *Bostonitic*—interlocking laths of feldspar, parallel in part and with crudely radial arrangement in other parts; bostonite, Boston, Massachusetts, U.S.A., XN.

Cenotypal—fine-grained fresh-looking and porphyritic, such as rocks of Tertiary or Holocene age.

6. *Clathrate*—leucite crystals with boundaries enhanced by the presence of acicular or stumpy pyroxene crystals oriented tangentially; leucitophyre with the pyroxene aegirine-augite, Eifel, France, PPL.

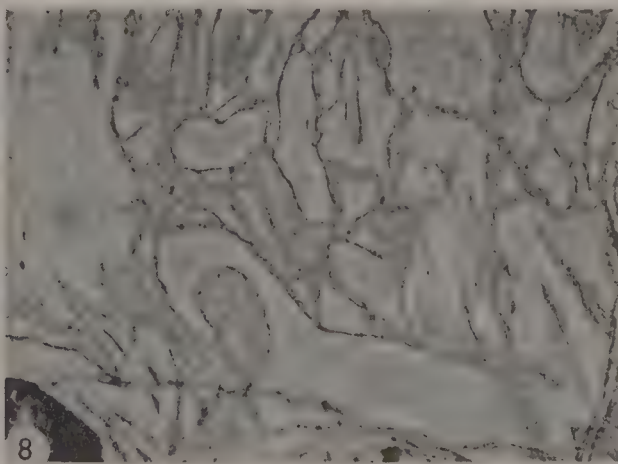
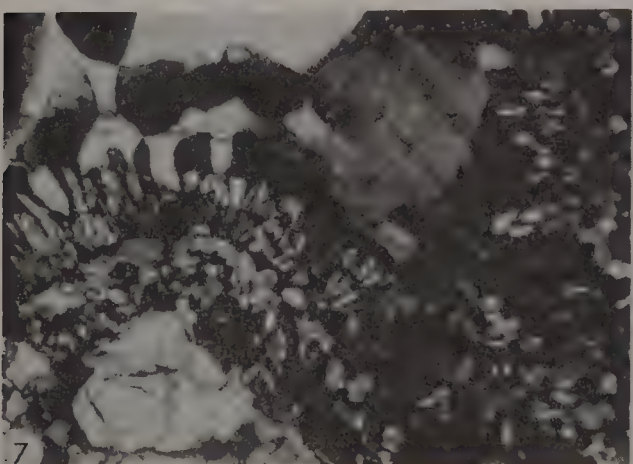
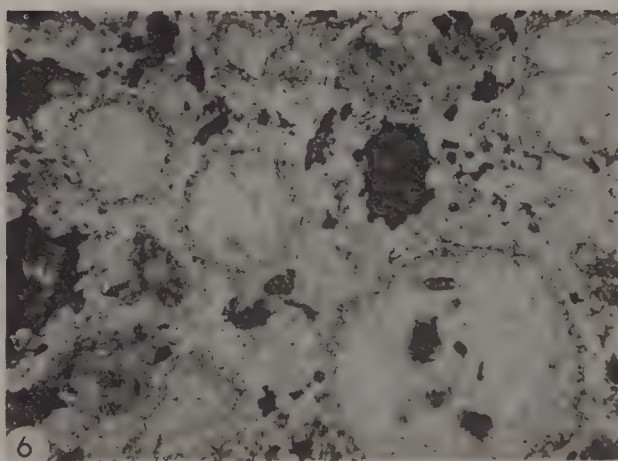
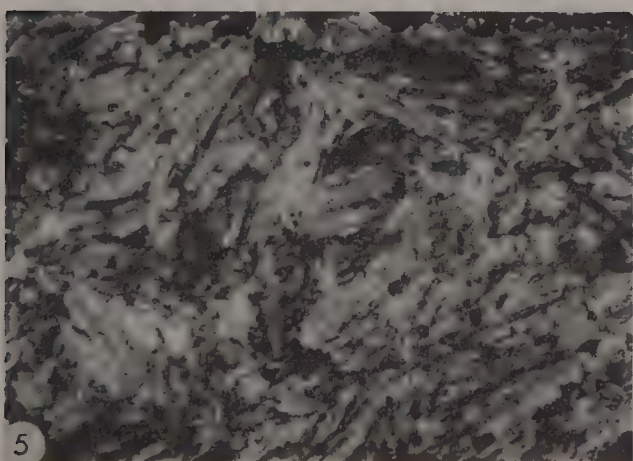
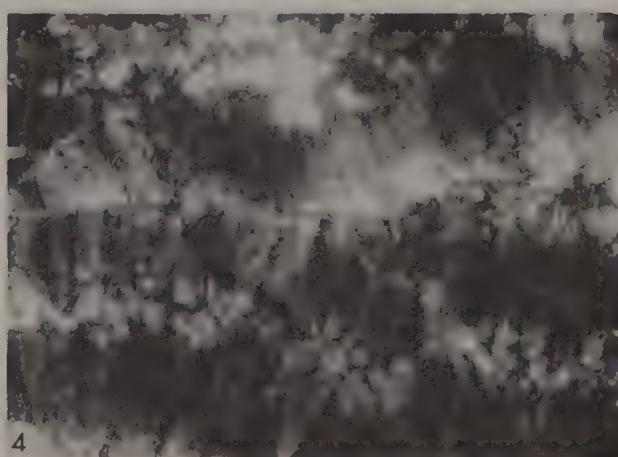
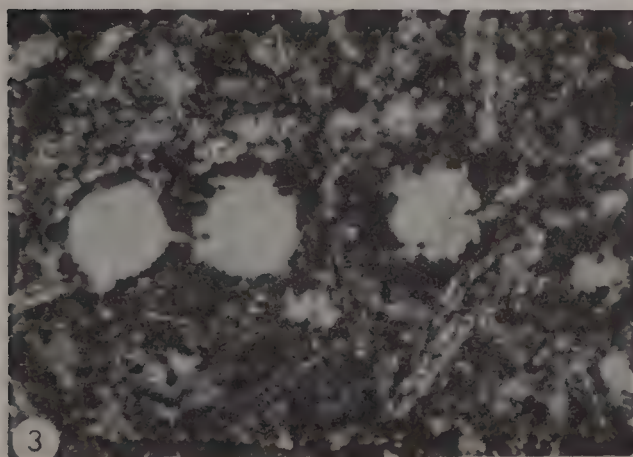
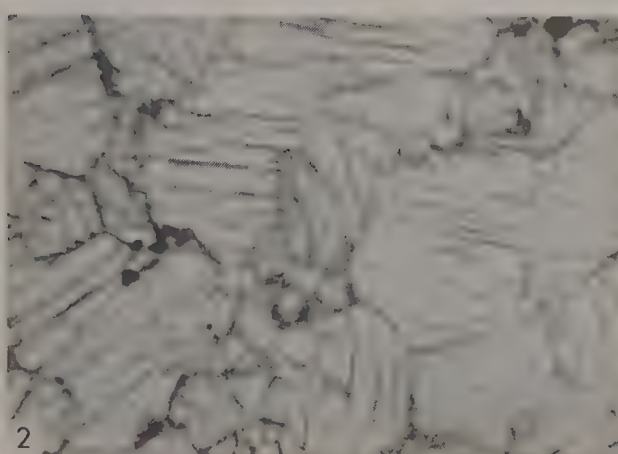
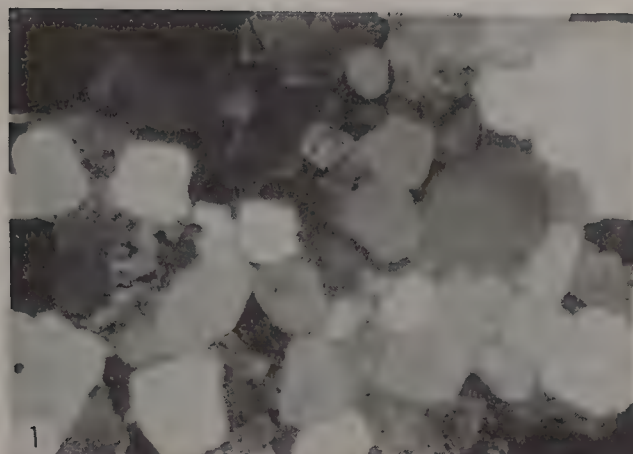
Consertal—irregularly shaped crystals closely interlocking without interstitial spaces.

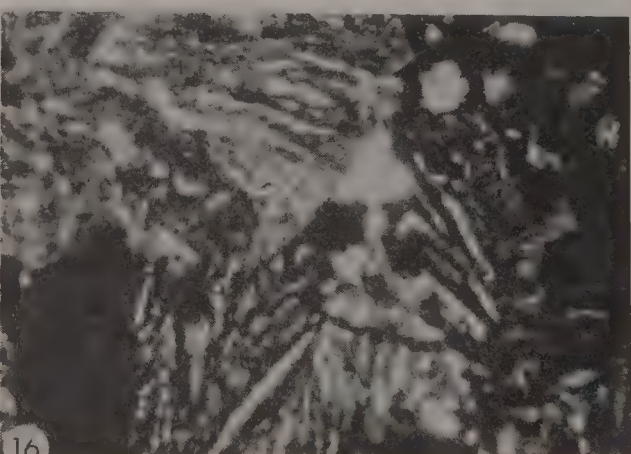
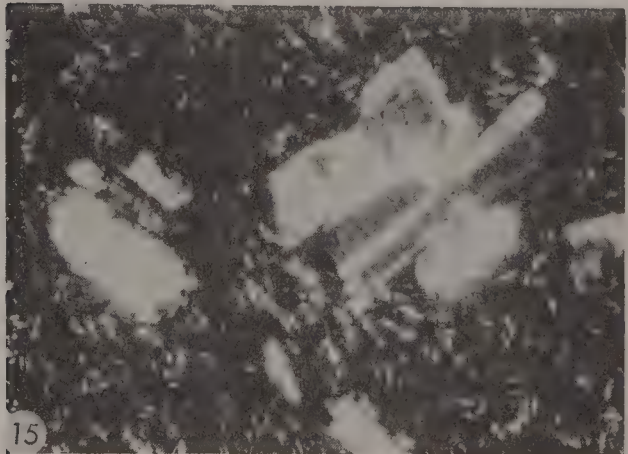
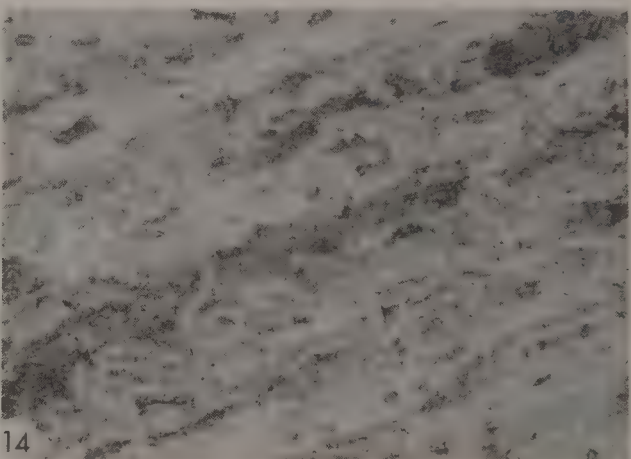
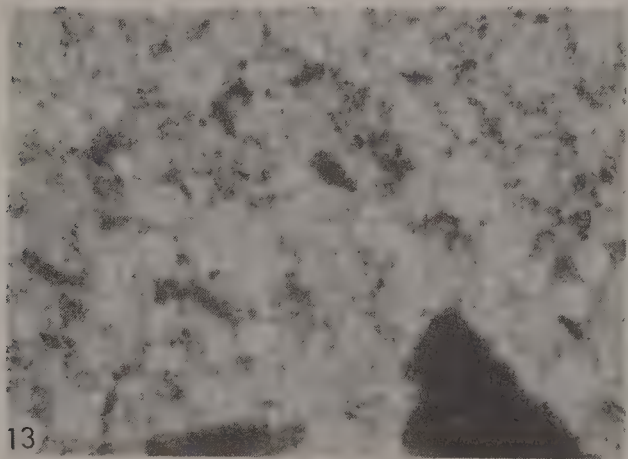
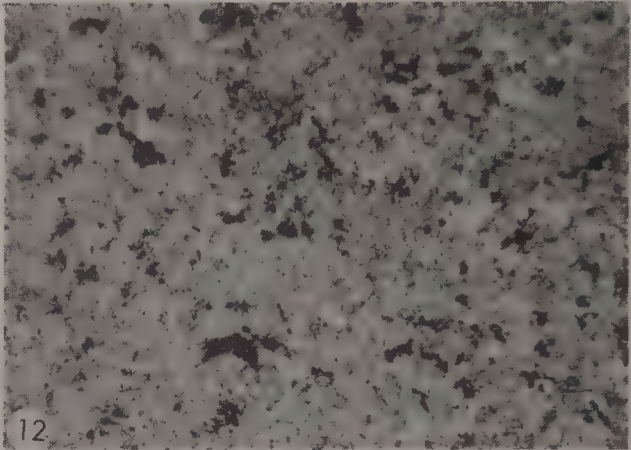
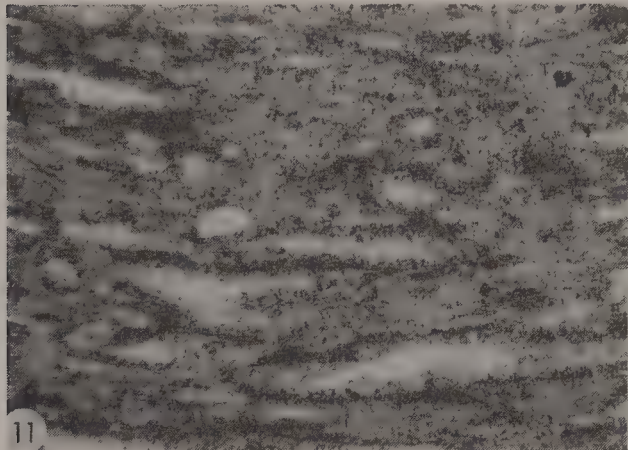
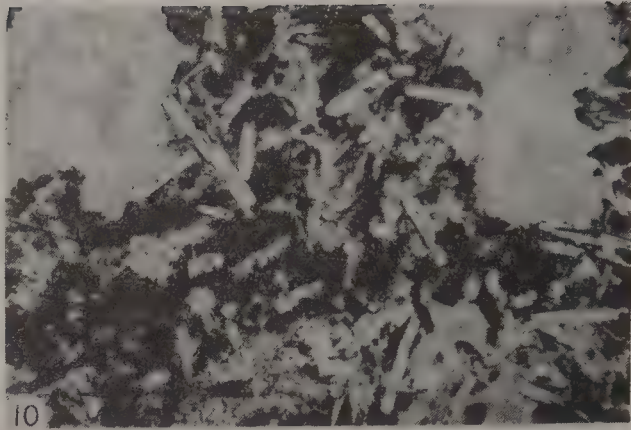
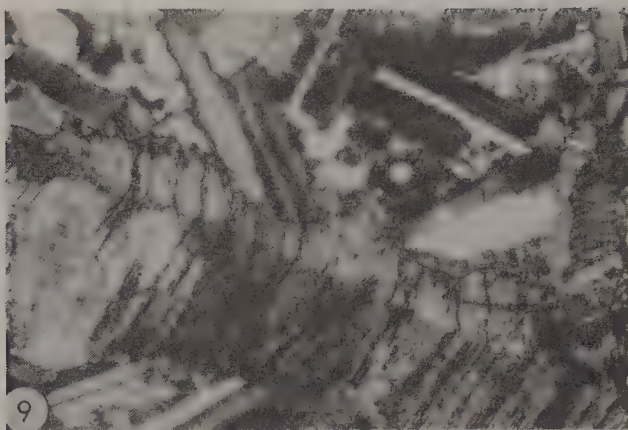
7. *Corona*—later member(s) of discontinuous reaction series surrounding earlier member(s); olivine surrounded by radially arranged stumpy prisms of pyroxene, intergrown with plagioclase; gabbro, Duluth, Minnesota, U.S.A., XN.

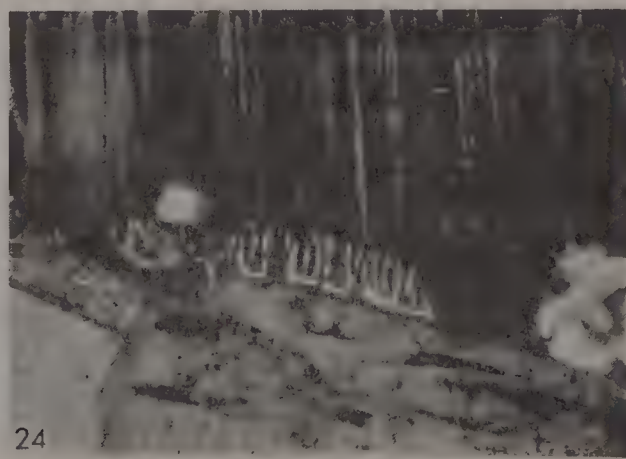
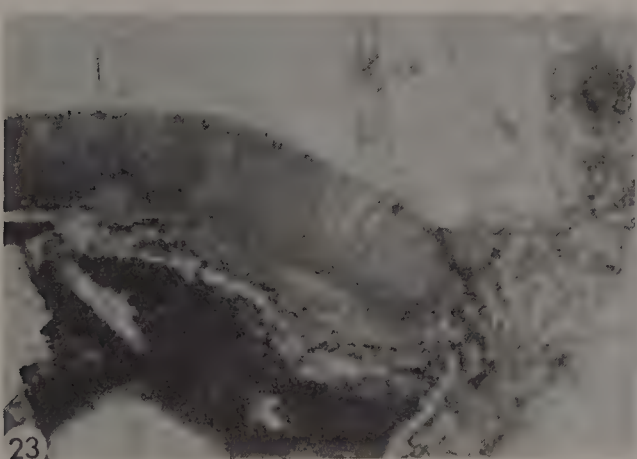
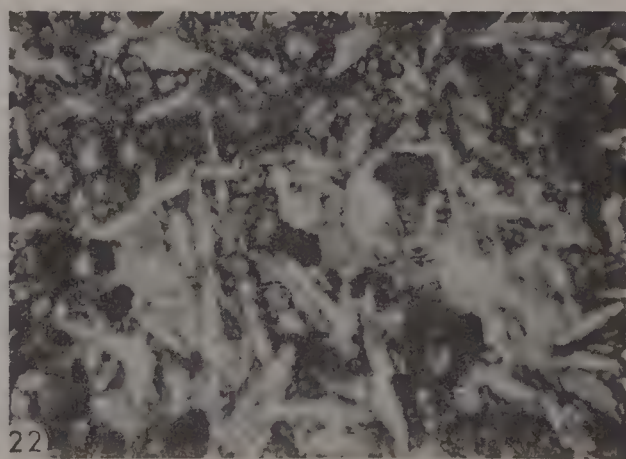
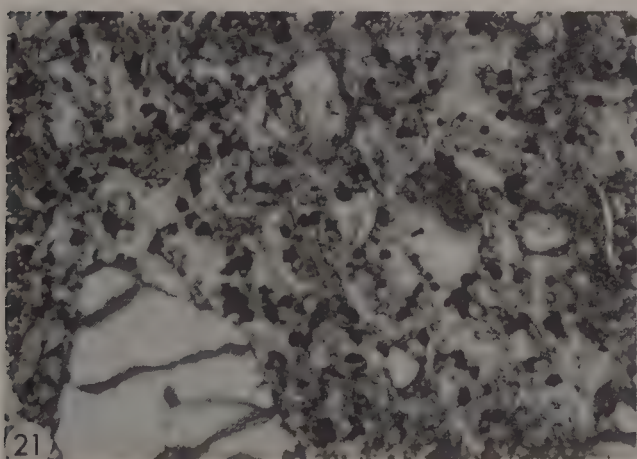
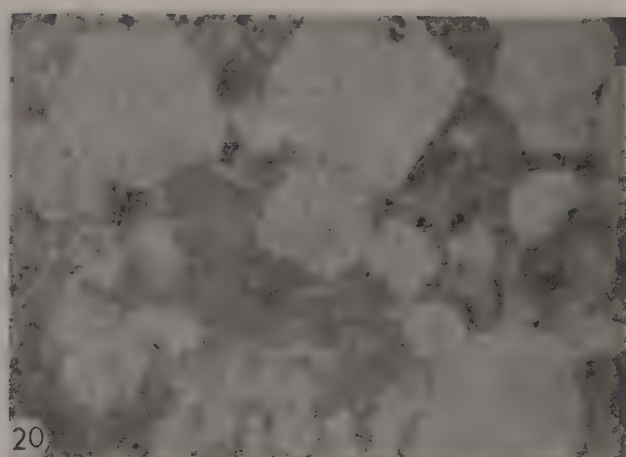
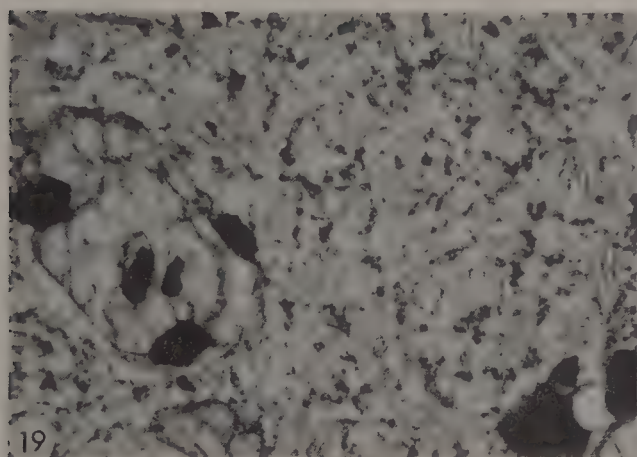
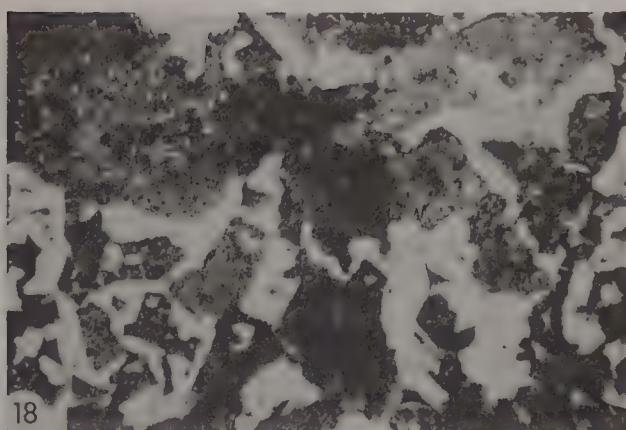
Cryptographic—intergrowth of quartz and feldspar on so fine a scale that individual minerals cannot be clearly distinguished under the microscope; a cryptocrystalline granophyric texture.

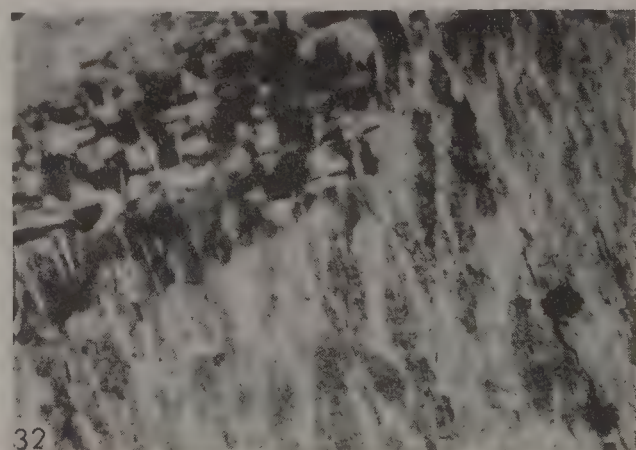
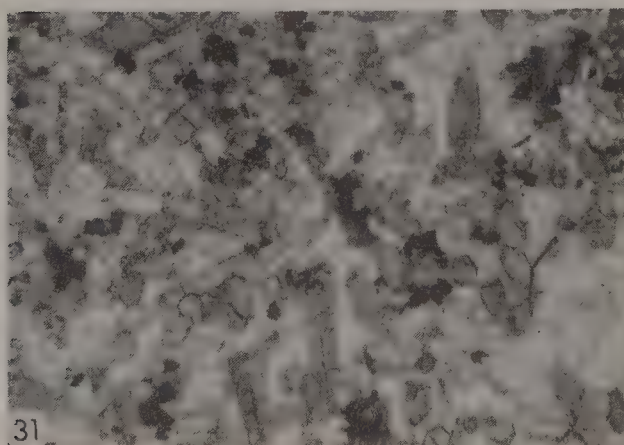
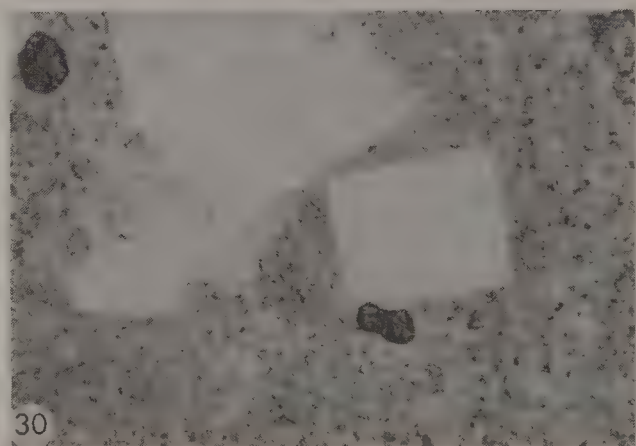
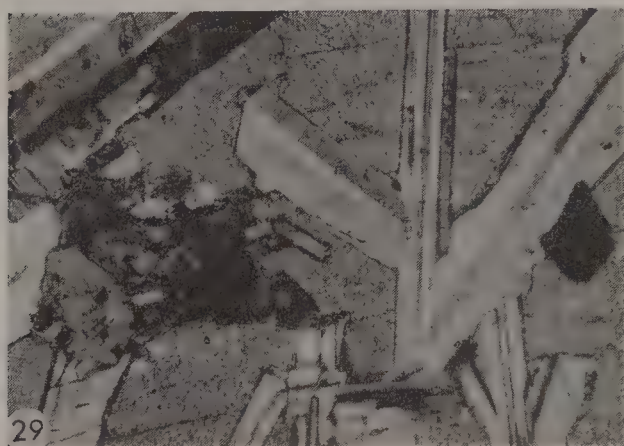
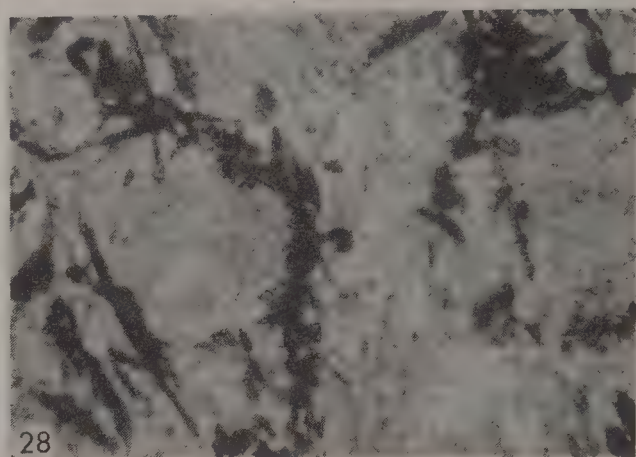
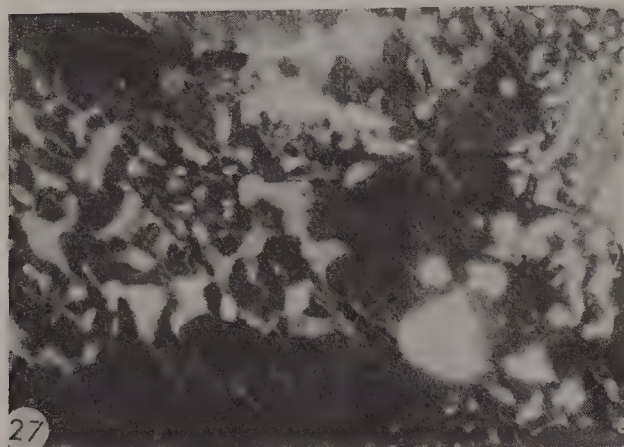
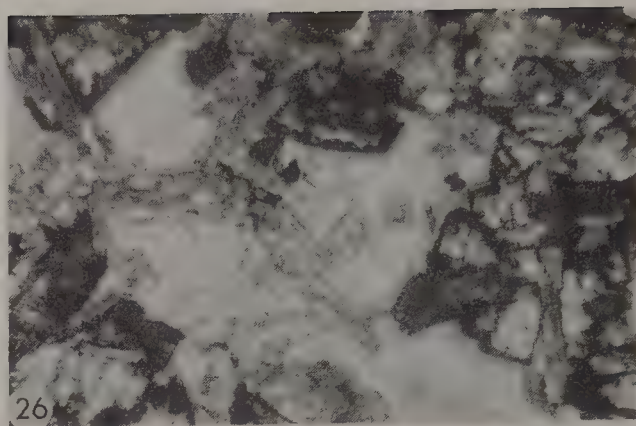
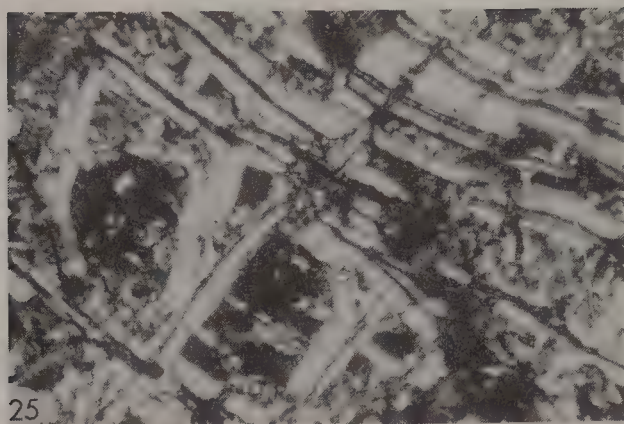
Crystallite—minute body of unknown mineralogical composition or crystal form that does not polarize light.

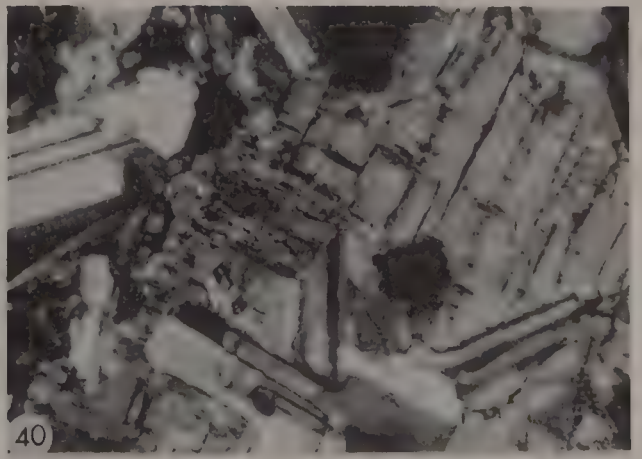
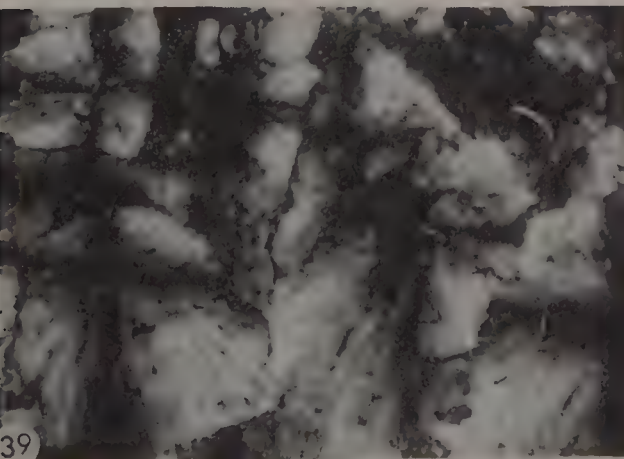
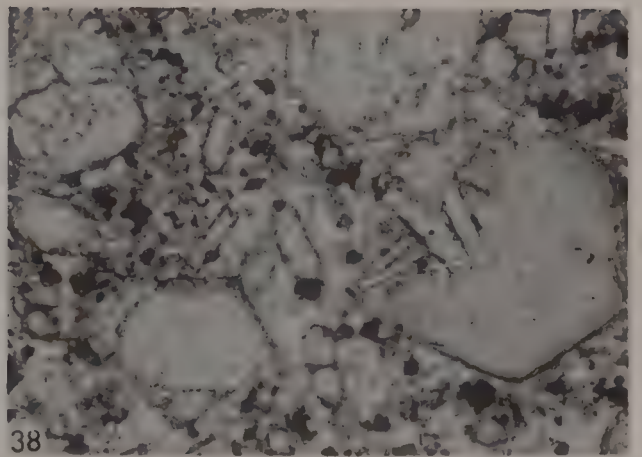
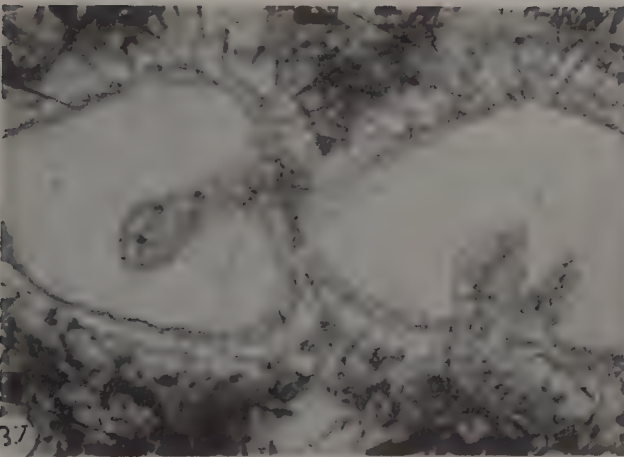
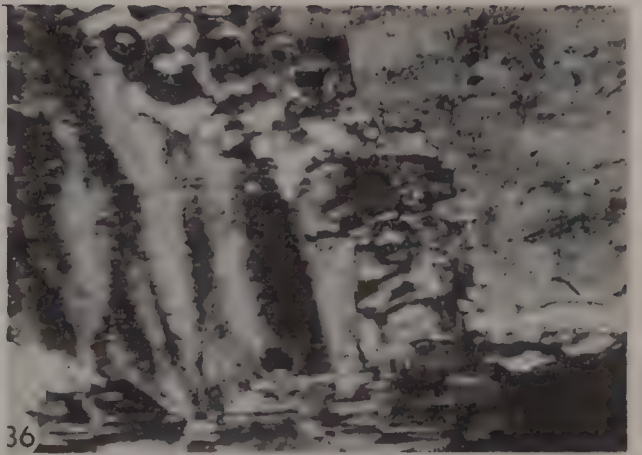
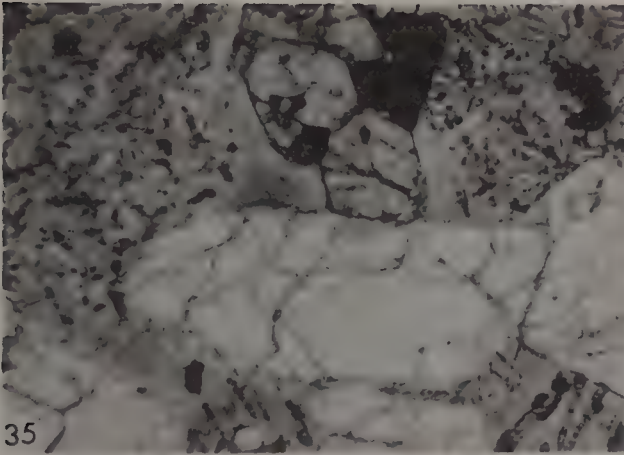
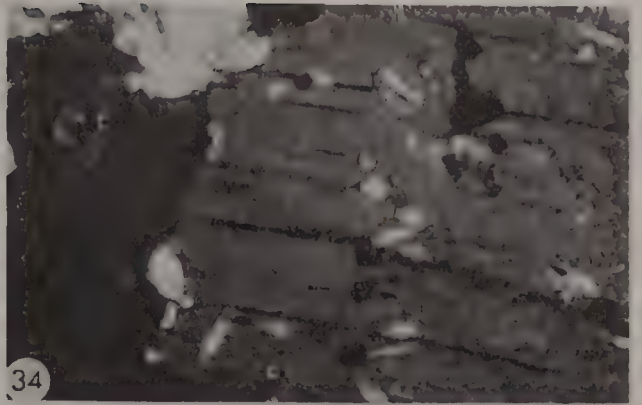
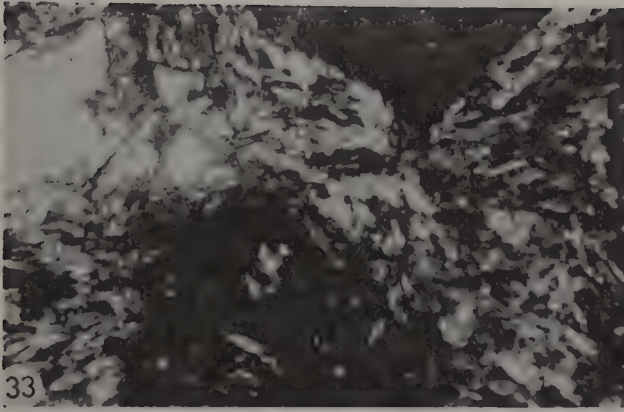
8. *Curlicue*—flattened glass fragments and shards resulting from flow and compaction; welded ash-flow tuff, Marysville, Utah, U.S.A., PPL, $\times 200$.

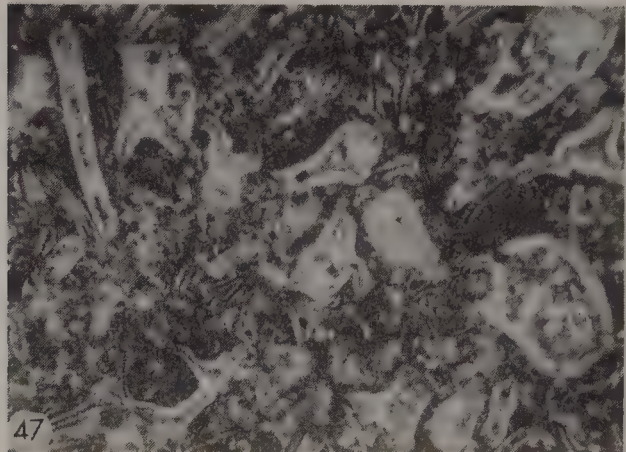
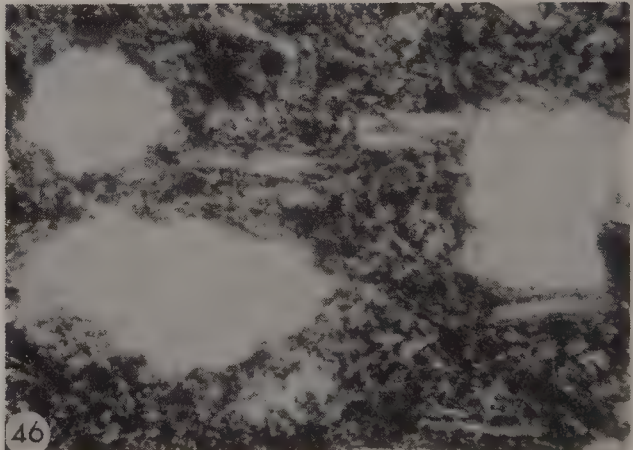
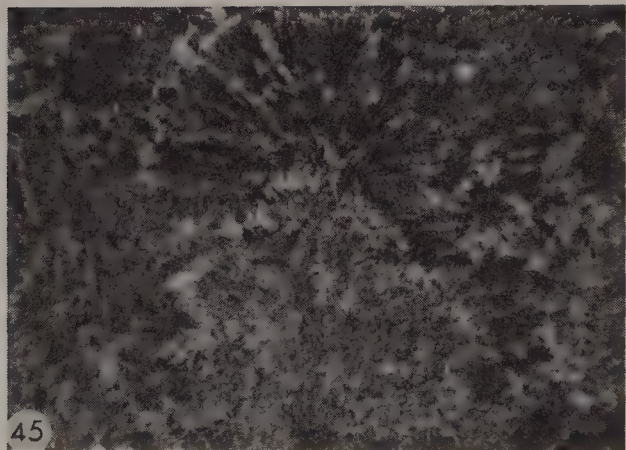
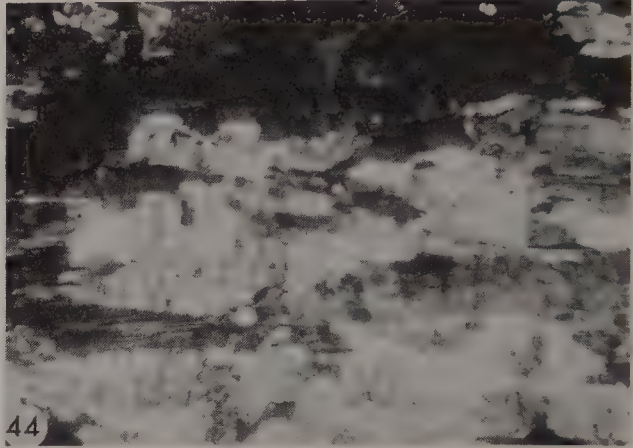
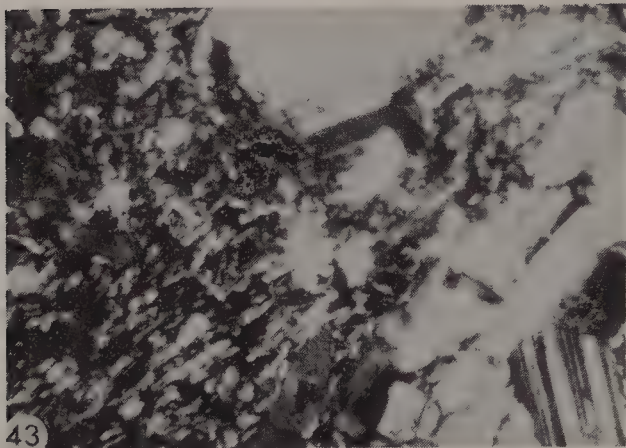
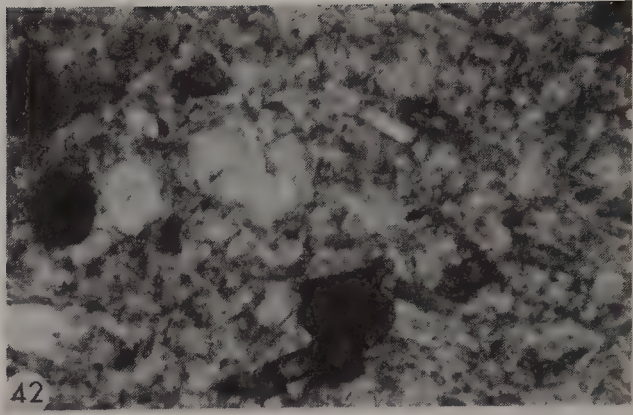
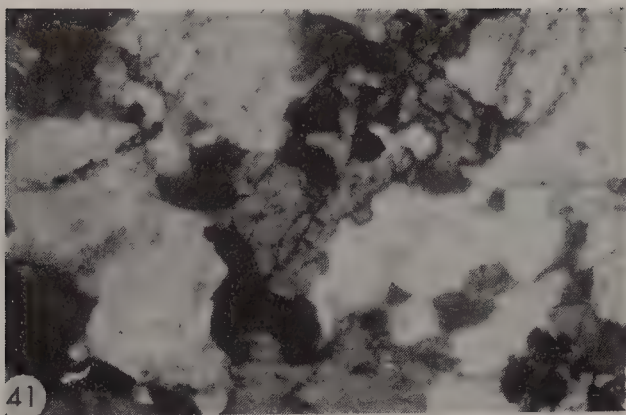












- Dactylitic*—symplectic intergrowth in which one mineral is penetrated by finger-like projections from another mineral (dactylitic intergrowth, dactylite).
9. *Decussate*—randomly criss-crossing, elongate, columnar laths or prismatic crystals—the term is more commonly used in relation to thermally metamorphosed rocks (hornfels) than igneous rocks; diorite with plagioclase laths and pyroxene showing *salite* texture (herringbone appearance), Salem, Massachusetts, U.S.A., PPL.
 10. *Diktytaxitic*—numerous jagged irregular small vesicles whose shapes are determined by the random orientation of small feldspar laths, some of which protrude into the cavities; basalt, Keweenaw Co., Michigan, U.S.A., PPL.
Druse (*drusy cavity*)—an irregular cavity, such as a steam hole in a lava, or in a vein, whose walls are encrusted with projecting minerals, generally the same minerals as those of the enclosing rock.
Embayment—a shape resulting from modification of a crystal or a xenocryst as the result of magmatic corrosion.
 11. *Eutaxitic (banded)*—blotches and streaks consisting of bands or elongate lenses that differ in color, composition, and texture and represent flattened and stretched rock fragments or devitrified glassy bands of different composition, particularly in welded ash-flow and ash-fall tuffs; welded tuff, Marysvale, Utah, U.S.A., PPL.
Feldsparphyric—contains feldspar phenocrysts.
Felsiphyric—microscopically aphanitic, or cryptocrystalline.
 12. *Felsitic (aphanitic)*—crystalline components in a dense rock not distinguishable by the unaided eye and positive identification commonly difficult or impossible; felsitic sometimes restricted to light-colored, quartz-feldspar rocks with aphanitic used for dark-colored rocks; felsite, Lake Superior region, Canada, PPL.
Felsitoid—mineral grains too small to be distinguished by the unaided eye.
 13. *Felsophyric (aphaniphyric)*—phenocrysts in a dense groundmass of intergrown quartz and feldspar; felsophyre, Augustusburg, W Germany, PPL.
Felty—microlites randomly oriented in pilotaxitic texture.
 14. *Flow (fluidal, rhyotaxitic)*—parallel aligned crystallites, color bands and/or fragments due to flow; effects of swirls and eddies due to turbulent flow may be shown; obsidian, Lake Co., Oregon, U.S.A., PPL.
Globulite—a spherical or globular crystallite in volcanic glass; cumulites are loosely aggregated cloudy masses of globulites; *globosphaerites* are more closely aggregated masses; *margarites* are linear strings of globulites.
Glomeroplasmatic—where granites or gneisses contain local open clusters of minerals such as biotite.
 15. *Glomeroporphyritic* (*cumulophyric-cumulophyre*)—crystal aggregates of one or more major components in a much finer-grained matrix; plagioclase porphyry, Boulder Co., Colorado, U.S.A., XN.
Graniphyric—phenocrysts in a microcrystalline groundmass.
Granitic—see *Hypidiomorphic granular*.
 16. *Granophyric*—irregular intergrowths of one mineral in another—commonly threads, patches and blebs of quartz in a base of feldspars; differs from micrographic (micropegmatitic) texture, which exhibits more regular cuneiform intergrowth; granophyre, Roskoqf, Vosges, France, XN.
Granular—aggregation of minerals of approximately equal size as in an equigranular holocrystalline rock.
Granulitic—granular with most crystals anhedral; the texture formed by crystals of augite and/or olivine occupying interstices of a network of plagioclase laths is termed *intergranular*.
 17. *Graphic, micrographic (pegmatitic,*
 18. *micropegmatitic)*—cuneiform intergrowths (quartz and feldspar in the case of pegmatitic) which resemble runic inscriptions; quartz-lepidolite pegmatite, Epprechtstein, Fichtelgebirge, W Germany, XN $\times 3(17)$; granite, Gavorrano, Italy, XN (18).
Gregaritic—phenocrysts in a groundmass in which independently oriented crystals occur in clusters.
Hyaline (hyalo-)—glassy.
Hyalinocrystalline—phenocrysts in a glassy groundmass.
Hyalocrystalline—phenocrysts in a glassy groundmass with proportions approximately equal (5:3–3:5).
Hyalo-ophitic—spaces in an open network of feldspar occupied by glass; resembles ophitic and is the limiting case of intersertal texture.
Hyalophyric—any porphyritic texture with a glassy groundmass.
 19. *Hyalopilitic*—an intersertal texture where the interspaces between feldspar microlites are filled with glass; glassy basalt (weiselbergite), Weiselberg, W Germany, PPL.
 20. *Hypidiomorphic-granular (hypoautomorphic-granular, granitic)*—generally even grained

with most essential components subhedral; some may be euhedral and some subhedral; granite, Weimduth, Massachusetts, U.S.A., PPL.

Idiomorph—a crystal bounded by most or all of its crystal faces (euhedral crystal); idiomorphic (*panidiomorphic* refers to the texture characterized by idiomorphs).

21. *Intergranular*—polygonal interstices between feldspar laths occupied by crystals of olivine, pyroxene and/or magnetite; basalt, Oelberg, Siebengebirge, W Germany, PPL.

22. *Intersertal*—similar to intergranular but with the interstices occupied by glass or cryptocrystalline material that may be of deuteric origin; basalt, Keweenaw Co., Michigan, U.S.A., PPL.

Isogranular—refers to the hypidiomorphic or idiomorphic texture of an igneous rock where the plagioclase crystals, and the pyroxene grains in the interstices between them, are of the same size.

23. *Kelyphitic (celyphytic) rim (border)*—borders or zones of cryptocrystalline material, or of parallel or radial associations of fibrous minerals, as the result of postmagmatic reactions between adjacent minerals; gabbro with kelyphitic rim between hypersthene and plagioclase, Duluth, Minnesota, U.S.A., PPL.

24. *Lacilinear*—worm-like inclusions of one mineral in the margins of another; granite showing quartz-feldspar intergrowths at the margins of perthite, Stearns Co., Minnesota, U.S.A., XN.

Linophyric—phenocrysts aligned in lines or streaks in a porphyritic rock (*linophyre*).

Lithoidal—individual constituents of a microcrystalline rock or groundmass, or a devitrified glass, too small to be distinguished by the unaided eye.

Margination—a texture shown by some granites characterized by sinuous contacts between quartz and feldspar grains resulting from magmatic corrosion.

25. *Mesh (reticulate)*—a network of zones of alteration with relict unaltered patches; serpentinite with serpentine developed from olivine, Silvaplana, Switzerland, XN.

Mesostasis—the last-formed interstitial material that is commonly glassy or aphanitic, but may be eutectic.

26. *Miarolitic*—drusy cavities of irregular shape with subhedral or anhedral crystals (commonly feldspars) projecting into them; these crystals are larger than in the main part of the rock; basalt, Keweenaw Co., Michigan, U.S.A., PPL.

Microlitic—composed of microscopic crystals (*microlites*, *microliths*) that polarize

light and have some determinable optical properties.

Miniphyric—phenocrysts with greatest dimension of <0.008 mm in a porphyritic rock (*mediiphyric* 0.008–0.04 mm, *magniphyric* 0.04–0.2 mm, *microphyric* <0.2 mm, *minophyric* 0.2–1 mm, *mediophyric* 1–5 mm, *magnophyric* >5 mm).

Monzonitic—euhedral plagioclase crystals with orthoclase occupying some of the interstitial spaces.

27. *Myrmekitic*—worm-like intergrowths (*myrmekite* refers only to quartz-plagioclase intergrowths); granite with quartz-orthoclase intergrowths in a perthite crystal, Gavorrano, Italy, XN.

Nesophitic—refers to ophitic texture in which pyroxene is interstitial to plagioclase and occurs in isolated areas.

28. *Ocellar*—similar to clathrate but with nepheline or analcime rather than leucite; an *ocellus* (pl. *ocelli*) is a phenocryst in a rock; phonolite with slender prismatic crystals of aegirine surrounding partly altered nepheline, dredged fragment, Goringe Bank, N Atlantic Ocean; PPL.

Oikocryst—see *Poikilitic*.

29. *Ophitic*—euhedral or subhedral laths of plagioclase enclosed in pyroxene crystals (typically augite); rarely referred to as *doleritic*; dolerite, New Jersey, U.S.A., XN.

30. *Orthophyric*—a variant of porphyritic with square-shaped phenocrysts, generally feldspars, in a fine-grained groundmass in which the feldspars have stumpy rectangular shapes rather than the slender lath-like shapes of many feldspar-bearing porphyritic rocks; rhyolite, Chaffey Co., Colorado, U.S.A., PPL.

Oxybasiophitic—refers to a rock with ophitic texture that may be either basiphitic or oxyophitic.

Oxyophitic—refers to a rock with ophitic texture whose mesostasis is composed of quartz and/or orthoclase; the term for the mesostasis is oxymesostasis.

31. *Panidiomorphic-granular (automorphic-granular, lamprophyric)*—generally even grained with most essential components euhedral; camptonite with barkevikite (dark) and plagioclase, Campton Falls, New Hampshire, U.S.A., PPL.

Perlitic—a system of irregular, convolute and spheroidal cracks in glassy material produced by contraction during cooling; may be found as a relict structure in devitrified rocks or in quartz and other minerals without cleavage.

32. *Perthitic*—intergrowth of K-feldspar and sodic plagioclase with shape and proportion of

blebs, blades or laminae of the enclosed mineral varying and the boundaries between host and inclusion generally ill-defined (perthite-plagioclase in K-feldspar host; antiperthite-K-feldspar in plagioclase host); granite with perthite, Rockport, Massachusetts, U.S.A., XN.

Phanerocrystalline (phaneritic)—all essential minerals can be distinguished individually by the unaided eye (*phanerite*) as distinct from aphanitic.

33. *Pilotaxitic*—lath-shaped microlites (usually plagioclase) arranged in a glass-free felty mesh, often aligned along flow lines (cf., trachytic); phonolite, Black Hills, S Dakota, U.S.A., XN.

Pipernoid—refers to the eutaxitic texture of some effusive rocks in which dark patches and stringers occur in a light-colored groundmass (*piperno*).

Planophyric—phenocrysts arranged in layers or laminae in a porphyritic rock (*planophyre*).

34. *Poikilitic (poecilitic)*—small crystals (chadacrysts) irregularly scattered and without common orientation in a larger host mineral (*oikocryst*) that exhibits optical and crystallographic continuity; when inclusions are numerous and round or subround in shape, the term *sieve* is used—more common in metamorphic than igneous rocks as a variant of poikiloblastic texture; gabbro with hornblende enclosing plagioclase and pyroxene, Salem, Massachusetts, U.S.A., PPL.

Poikilophytic—refers to ophitic texture characterized by lath-shaped feldspar crystals.

- 15, *Porphyritic*—large crystals (*phenocrysts*), 33, often euhedral, in a *groundmass (matrix, 36, mesostasis)* of either much smaller crystals or 45. glass; where the crystals are markedly different in size, the term *hiatal* can be used.

Porphyrogranulitic—refers to an ophitic texture characterized by large phenocrysts of feldspar and augite and/or olivine in a groundmass of lath-shaped feldspar crystals and irregular augite grains; a combination of porphyritic and intergranular texture.

35. *Protoclastic*—earlier-formed minerals, such as phenocrysts, but not the groundmass minerals, show fracturing and/or marginal granulation indicative of differential flow before complete consolidation; weiselbergite with much fractured plagioclase, Weiselberg, W Germany, PPL.

Radiolite—radial fan-like groupings of acicular crystals that resemble sectors of spherulites.

36. *Rapakivi*—ovoidal K-feldspar phenocrysts rimmed (mantled) by sodic plagioclase that may be as minute granules with random

orientation; granite, Rockport, Massachusetts, U.S.A., XN.

37. *Reaction rim*—a peripheral zone of one mineral around another, commonly due to reaction with the enclosing melt (*corona*), but also as the result of postmagmatic reaction (*kelyphitic rim*), late or postmagmatic alteration by residual magmatic fluids, or magmatic reaction with a xenocryst; pitchstone with embayed quartz surrounded by radially arranged feldspar, Hlinik, Hungary, PPL.

Reticulate—crystals partly altered to a secondary mineral form a network enclosing remnants of the original mineral.

Rhyodiabasic—refers to an ophitic texture in which plagioclase phenocrysts are more or less parallel.

Scopulite—a crystallite the shape of a rod or stem that terminates in brushes or plumes.

Scoriaceous (scoriform, scorious)—refers to the texture of a coarsely vesicular rock, usually andesitic or basaltic in composition; vesicle walls may be either smooth or jagged.

38. *Seriate*—sizes of crystals vary gradually or in a continuous series; vogesite with hornblende crystals of different sizes, Four Corners, Colorado Plateau, U.S.A., PPL.

Serrate—saw-tooth contacts between minerals usually due to replacement.

Skedophyric—phenocrysts more or less uniformly distributed throughout the groundmass of a porphyritic rock (*skedophyre*).

39. *Spherulitic*—numerous spherical or subspherical bodies (*spherulites, sphaerolites*) with radial arrangement of fibrous or acicular crystals, commonly feldspar, around a central grain or empty cavity; glassy rhyolite, Marysvale, Utah, U.S.A., XN.

Spheruloid—a spherule or nodule without radial structure as in perlitic lava.

Spiculite—a spindle-shaped belonite.

Spinifex—skeletal, platy, or acicular crystals of olivine, orthopyroxene, or clinopyroxene, or their pseudomorphs, forming a pattern that, on the weathered surface of a rock, looks like the Australian desert grass *Triodia spinefex*; in Canada the term is synonymous with “chicken-track,” “bird’s foot,” “bird-trap,” “herring-bone,” “feather,” or “lath” texture (see **Spinifex texture**).

Subdiabasic—similar to ophitic texture except that the augite is not optically continuous but present as granular aggregates.

40. *Subophitic*—ophitic texture with the feldspar crystals approximately the same size as the pyroxene and only partly included by

them; dolerite, Suffolk Co., Massachusetts, U.S.A., PPL.

41. *Sutured*—interlocking, suture-like mineral boundaries—more common in metamorphic than igneous rocks; nepheline syenite with microcline and arfvedsonite, Wausau, Wisconsin, U.S.A., XN.

Symplectic—intimate intergrowth of two minerals as in graphic, micrographic, pegmatitic, and micropegmatitic, but sometimes restricted to minerals of secondary origin (*symplectite*).

Synantectic—refers to a primary mineral formed by the reaction of two other minerals, such as in the development of a reaction rim.

Synneusis—refers to a texture of a rock in which some of the crystals are aggregated in clusters; it is not restricted to a porphyritic rock nor does it have the mineral clusters form the phenocrysts as in glomeroporphyritic.

42. *Tinguaitic*—a porphyritic texture in which the groundmass is made up of small equant crystals surrounded by mafic acicular crystals arranged in radial or criss-cross patterns; tinguaite with nepheline phenocrysts and aegirine needles in groundmass, Oberburgen, W Germany, PPL.

Trachyophitic—refers to an ophitic texture in which the feldspar crystals enclosed by pyroxene have a parallel or subparallel alignment.

43. *Trachytic*—parallel arrangement of feldspar laths that wind around phenocrysts; trachyte, Drachenfels, Siebengebirge, W Germany, XN.

44. *Trachytoid*—feldspar laths or elongate crystals in parallel arrangement in medium- and coarse-grained rocks; syenite, Wausau, Wisconsin, U.S.A., XN.

Trichite—straight or curved hair-like crystallite, usually black and occurring singly or in radially arranged clusters in glassy igneous rocks.

45. *Variolitic*—pea-sized spherical bodies (*varioles*) showing radiant or divergent arrangement of plagioclase fibers; basalt, Quelle de Durance, France, PPL.

46. *Vesicular*—abundance of *vesicles* (*vacuoles*) which may be spherical or elongate because of compaction or flow; glassy rhyolite, Marysvale, Utah, U.S.A., XN.

47. *Vitroclastic*—glass shards and glass fragments of crescentic or roughly triangular outlines derived from fragmentation of vesicular material; rhyolitic tuff, Marysvale, Utah, U.S.A., PPL.

48. *Vitrophyric* (*porphyritic glassy*)—phenocrysts or crystal (and lithic) fragments in a

glassy matrix; rhyolite with perlitic cracks, Marysvale, Utah, U.S.A., PPL.

ANNA T. GAVASCI

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TEXTURES OF METAMORPHIC ROCKS

The following are textures, or terms used in relation to textures, of metamorphic rocks; * indicates an illustration in the cross-reference.

Allothimorph—a constituent of a metamorphic rock that has not changed its original crystal outline.

Allotrioblast(ic)—see **Xenoblast(ic)**

Atoll—see ***Atoll garnet**.

Augen—an eye-shaped large crystal or mineral aggregate (German *auge*, eye) in gneiss or schist; it originates by (1) flow banding of phenocrysts in a syntectonic pluton; (2) the growth of porphyroblasts in a banded rock (*augen-blast*; cf., blastomylonite); (3) the cataclastic reduction of large crystals (*augen-blast*); and (4) the cataclastic reduction of brecciated fragments (*augen-clast*); the dimensional alignment of augen in gneisses or schists defines an *augen structure* (see ***Gneiss**; ***Porphyroclast**).

Crystalloblast—a crystal of a mineral produced entirely by metamorphic processes; the texture produced by metamorphic recrystallization is referred to as *crystalloblastic texture*; *crystalloblastic series* is an arrangement of metamorphic minerals in order of decreasing form energy so that a crystal of a mineral in a high position in the series tends to assume idioblastic outlines at surface contact with simultaneously developed crystals of all minerals occupying lower positions in the series.

Decussate—a texture of thermally metamorphosed rocks (e.g., hornfels) characterized by the cross-cross, diverse arrangement of minerals

as a mechanical expedient for minimizing internal stress (see **Hornfelsic textures**).

Deutomorphic—refers to crystals whose shapes have been acquired or modified by the action of mechanical or chemical processes acting on the original forms.

Diablastic—a texture shown by intricately ingrown and interpenetrating minerals, usually with rod-like shapes.

Fibroblastic—a variant of homeoblastic texture with minerals of fibrous habit developed during recrystallization.

Flaser—lenses and layers (flaser) of original or little-altered granular material surrounded by a matrix of highly deformed material (giving the appearance of a crude flow structure) make up flaser structure as in flaser gabbro (see **Cataclastic rocks**; ***Gneiss**).

Foliation—planar arrangement of dimensionally oriented minerals and mineral bands as in gneiss; also used for schistosity in a schist and cleavage in slate.

Glomeroplastic—texture of gneisses (or granites) containing local open clusters of certain minerals, such as biotite.

Gneissose (gneissic)—a texture (or structure) made up of an alternation of (schistose) bands and lenticles of dimensionally oriented flaky or elongate prismatic and (granulose) bands of granular minerals, e.g., hornblende and/or biotite with quartz-feldspar (***Gneiss**).

Granoblastic—a variant of homeoblastic texture resulting from the development of essentially equidimensional crystals, generally with sutured boundaries.

Granoschistose—refers to the structure of a monomineralic rock produced by the parallel elongation of grains of a mineral that is normally equidimensional, or nearly so, as in a quartz schist.

Granuloblastic (granoblastic polygonal)—a texture due to a mosaic of xenoblasts with smooth intergranular boundaries or straight edges approximating to a polygonal arrangement, as in many rocks of the granulite facies.

Granulose—a texture (structure) characterized by the presence of granular minerals such as quartz, feldspar, garnet, and pyroxene in granulites; alternating streaks or bands may vary in mineralogy, but consist of granular minerals.

Helicitic (helizitic)—a texture consisting of bands of inclusions indicative of a previous feature (schistosity, bedding) in later-formed crystals, particularly porphyroblasts; the fabric within the porphyroblast is referred to as Si (internal) and a new planar fabric in the rock as Se (external) (see ***Polyphase metamorphic mineral growth**).

Heteroblastic—a variant of crystalloblastic

texture in which the essential minerals are of two or more distinct orders of magnitude of size.

Holoblast—a crystalloblast entirely formed during metamorphism.

Homeoblastic—a variant of crystalloblastic texture in which the essential minerals are of approximately equal size (cf., **Fibroblastic**; **Granoblastic**; **Lepidoblastic**; **Nematoblastic**).

Hornfelsic—see ***Hornfelsic textures**.

Hypidioblast(ic)—a crystalloblast bounded only by some of its characteristic crystal faces (analogous to hypidiomorphic or subhedral in igneous rocks).

Idioblast(ic)—a crystalloblast bounded by its characteristic crystal faces (analogous to idiomorphic or euhedral in igneous rocks)—see ***Porphyroblast**.

Idiotopic—a special case of granoblastic texture in which most of the crystals are bounded by crystal faces.

Lepidoblastic—a variant of homeoblastic texture exhibited by schistose (foliated) rocks resulting from the presence of minerals with a flaky or scaly habit (see ***Polyphase metamorphic mineral growth**).

Maculose—the texture of spotted or knotted slates, etc., formed by contact (thermal) metamorphism.

Mortar—a texture (structure) of cataclastic rocks consisting of relics (often eye-shaped with unbroken cores but recrystallized terminations) surrounded by an aggregate of finer-grained, crushed and recrystallized grains, commonly mica-free, of the same minerals as the relics (see **Cataclastic rocks**; ***Gneiss**; **Porphyroblast**).

Mosaic—a variant of granoblastic (or granoblastic polygonal) texture in which the individual grains meet with straight or only slightly curved, but not interlocking boundaries.

Murbruk—see **Mortar**.

Mylonite—a streaky or banded structure that somewhat resembles flow structure resulting from intense ductile deformation; characteristic of mylonites (see **Cataclastic rocks**).

Nematoblastic—a variant of homeoblastic texture characterized by the presence of slender prismatic crystals (see ***Polyphase metamorphic mineral growth**).

Palimpsest—a structure or texture in which remnants of a previous structure or texture are preserved.

Phacoidal—a structure characterized by the presence of lenticular crystals or crystal aggregates (e.g., augen structure, flaser structure).

Plättung—a texture characterized by much-flattened quartz grains resulting from intense ductile deformation, particularly in granulite facies environments (see **Granulite**).

Poikiloblast(ic), *poeciloblast(ic)*—a large crystal formed by metamorphic recrystallization that contains numerous small inclusions (see *Polyphase metamorphic mineral growth; Retrograde metamorphism; Schist).

Porphyroblast(ic)—see *Porphyroblast.

Porphyroclast(ic)—see *Porphyroclast.

Pseudophenocryst, *pseudoporphyrityc*—see **Porphyroblast**.

Pseudoporphyrityc—relates to a structure that resembles porphyroblastic texture, but is due to processes other than prograde crystal growth, e.g., differential granulation.

Relic (noun), *relict* (adjective)—relates to a mineral, structure, or other feature of a rock that persists during metamorphic reconstitution (stable relic, unstable relic)—see *Palimpsest*.

Schistose—a texture characterized by the parallel, planar arrangement of mineral grains of platy, prismatic, or ellipsoidal types, particularly micas, that makes up schistosity (see *Schist).

Sieve—a texture that is a particular expression of poikiloblastic texture in which there is a regular distribution of inclusions throughout the poikiloblast.

Snowball—see *Snowball garnet.

Sutured—a texture resulting from mineral grains or irregularly shaped crystals interfering with adjacent grains and producing interlocking, irregular contacts without interstitial spaces, resembling sutural structures in bones; also said of crystal contacts in a rock showing these features; mainly shown in metamorphic rocks, but also shown by some igneous rocks.

Xenoblast(ic), *allotrioblast(ic)*—a crystalloblast that does not show its characteristic faces and the resultant texture.

D. R. BOWES

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THERMAL ACTIVITY

Thermal activity refers to individual or groups of thermal manifestations on the Earth's surface, such as hot springs, hot ground, geysers, fumaroles, and solfataras. These phenomena are the surface outlets of hydrothermal systems involving the circulation of groundwater to considerable depths. Three basic types of thermal systems have been recognized: (1) low-temperature systems involving subsurface temperatures either below or not much in excess of 100°C and a fluid phase

throughout; (2) wet high-temperature systems with subsurface temperatures considerably above 100°C but containing a mixed phase, that is, steam and water; (3) vapor-dominated high-temperature systems with high subsurface temperatures in which the steam phase prevails throughout the system.

The high-temperature systems are usually quite extensive. Some of the larger systems generate surface displays scattered over areas more than 100 km². The circulation may extend down to depths of several kilometers, or more. Observed temperatures at depth are 200°–380°C. The total rate of heat escape from the ground in the case of the largest individual high-temperature areas may amount to 10⁸–10⁹ J/sec. The total heat output of all thermal activity on the Earth's surface appears to be of the order of 10¹⁰ J/sec., i.e., about a thousandth part of the total rate of outward conduction of terrestrial heat.

High-temperature activity is invariably connected with recent volcanism. Many of the largest high-temperature areas are located within the circum-Pacific volcanic belt, i.e., in New Zealand, Indonesia, Japan, Alaska, western U.S.A., Central America and in the Andes, S America. Moreover, very active areas of this type are located in Iceland, Italy, Ethiopia, Kenya, and in the W Indies.

Low-temperature activity is also present in many volcanic regions. However, areas of this type are also found in regions where volcanism has not been active for several geologic periods, e.g., Warm Springs, Georgia, U.S.A.

There is little doubt that high-temperature thermal activity is maintained by heat escaping from magmatic intrusions at relatively shallow depth. At least some of these areas are clearly postvolcanic phenomena. Low-temperature activity in nonvolcanic regions, on the other hand, may be the result of a very deep circulation of water.

Studies of the energy balance of thermal areas and studies of the isotopic composition of the water issued by thermal springs have shown that most, if not all of the water is of meteoric origin. Only high-temperature postvolcanic fumaroles or solfataras of very recent origin may issue appreciable quantities of juvenile water, that is, water that has been carried up in solution from great depths by ascending magma.

G. BODVARSSON

For literature references see cross-referenced articles.

Cross-references: *Fumarole*; *Geyser*; *Hot ground*; *Hot spring*; *Solfatara*; *Volcanoes and volcanism*.

TRACHYANDESITE, TRACHYBASALT

Trachyandesite is an effusive igneous rock intermediate in composition between trachyte and andesite, composed of plagioclase (andesine–oligoclase), K-feldspar, and one or more mafic minerals (biotite, amphibole, pyroxene). *Trachybasalt* is an effusive igneous rock intermediate in composition between trachyte and basalt, composed of calcic plagioclase, K-feldspar (commonly sanidine) with augite, olivine, and possibly minor analcime or leucite. Both rock types have been variously equated with one another and with latite; also, trachyandesites have been equated to latites plus latite-andesites and trachybasalt to latite and latite-basalt. *Trachydolerite*, an alkali basalt with orthoclase or anorthoclase, labradorite, and a small proportion of feldspathoids, and a chemical composition intermediate between trachyte and basalt, has also been used as a synonym of trachybasalt.

Rocks on a direct compositional line from (alkalic) basalt to (sodic) trachyte are hawaiiite, mugearite, and benmoreite, whereas trachybasalts tend to be in the differentiation series of alkali basalt → trachybasalt → tristanite → potassic trachyte. The generally Q-normative trachyandesites lie on the tholeiite → trachyandesite → trachyte → rhyolite trend.

The following are various terms used in connection with this rock association.

Arso trachyte—an olivine trachyandesite with phenocrysts of sanidine, oligoclase, augite, and olivine in a trachytic groundmass containing interstitial glass; related to kenyte and rhombporphyry.

Benmoreite—an alkaline rock intermediate between mugearite and trachyte and for which proportions of normative minerals calculated from a chemical analysis is commonly needed for precise identification; differentiation index 65–75 and $K_2O:Na_2O < 1:2$.

Dancalite—an analcime trachyandesite with phenocrysts of oligoclase and augite in a trachytic groundmass containing analcime.

Dumalite—a trachyandesite with intersertal texture; the glassy mesostasis possibly has the composition of nepheline.

Gauteite—a porphyritic hypabyssal rock with the composition of a trachyandesite; its bostonitic groundmass is composed mainly of andesine (Gaute, Czechoslovakia).

Isenite—a trachyandesite with phenocrysts of andesine, Na-microcline, hornblende, and biotite and a groundmass of oligoclase, orthoclase, and nosean with augite, apatite and opaque ores.

Mugearite—a dark-colored, fine-grained rock with oligoclase the main feldspar and subordinate orthoclase; olivine and Fe–Ti oxides are more abundant than pyroxene; intermediate between hawaiiite and benmoreite with DI 45–65 (Mugeary, Skye, Scotland).

Parchettite—a mesocratic leucite trachyandesite.

Sancyite—a tridymite-bearing leucotrachyandesite.

Tahitite—a trachyandesite with phenocrysts of h         and generally more sodic plagioclase than orthoclase (Tahiti).

Trachylabradorite—a trachybasalt with plagioclase $> An_{50}$.

Tristanite—rock intermediate between trachyandesite and trachyte: differentiation index 65–75 and $K_2O:Na_2O > 1:2$ (cf., *Benmoreite*) (Tristan de Cunha).

D. R. BOWES

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Cross-references: *Andesite and dacite*; *Basalt*; *Igneous rocks—classification and nomenclature*; *Intermediate igneous rocks*; *Trachyte*; *Volcanic rocks*.

TRACE ELEMENTS: GEOCHEMISTRY—See Vol. IVA

TRACHYTE

Trachytes are extrusive or high-level intermediate igneous rocks, composed dominantly of fine-grained alkali-feldspar usually with phenocrysts of ternary anorthoclase feldspar or of coexisting sanidine and sodic plagioclase. Potassic trachytes (*sanidinites*) occur rarely. With increasing plagioclase, trachytes grade into trachyandesite (latite of some authors) and andesite; with increasing quartz they grade into rhyolites. Undersaturated trachytes with minor groundmass feldspathoid occur and ultimately grade into phonolite. Tabular matrix feldspars in parallel orientation, a result of magmatic flow, are very common and give rise to the so-called trachytic texture. Groundmass glass may be present, and trachytes may be vesicular and sometimes agglomeratic. The name is due to Ha       (1822—see Johannsen, 1937), and is an illusion to the surface roughness of many trachytes.

Mineralogy

The feldspar phenocrysts may be large and sporadic, as in the celebrated Drachenfels trachyte, so that some care is required in using thin-sections to decide the type of feldspar present. In one-feldspar trachytes, phenocrysts are ternary anorthoclase feldspars, often cryptoperthitic. In sanidinites, partially resorbed plagioclase may occur as cores to sanidine phenocrysts. Two-feldspar trachytes contain a sodic plagioclase (sometimes repeatedly and strikingly zoned) that may be andesine or oligoclase, together with Na-bearing sanidine, also often cryptoperthitic. The nature of the feldspar phenocrysts essentially reflects the normative *An* content of the melt; compositions of feldspar pairs in trachytes have been discussed by Carmichael (1965—and see **Syenite**). Groundmass feldspar is usually largely alkali feldspar laths, but some plagioclase may be present.

Mafic phases (as small and rarely abundant phenocrysts, or in the matrix) indicate the peralkaline tendencies of the trachyte. Amphibole may be green or brown hastingsite grading to alkali amphibole (arfvedsonite); pyroxenes range from diopside to ferroaugite or sodic ferroaugite. Amphibole and biotite are often partially or completely resorbed and replaced by iron-ore. Alkaline trachytes may contain fayalitic olivine. Sphene and apatite are common accessory minerals. Quartz (or the high-temperature polymorphs cristobalite and tridymite) may occur, and undersaturated trachytes may carry nepheline or sodalite as a minor constituent.

Distribution and genesis

Trachytes are comparatively rare but persistent members of oceanic basaltic suites, and may be extruded with rhyolites by the same volcano. They are found most abundantly in tracts of undersaturated rocks such as the E African rift system (Clifford and Gass, 1970) or in more ancient alkaline provinces such as the Gardar province in S Greenland where trachytic hypabyssal types occur (Emeleus and Upton, 1976). In such provinces trachyte, or its plutonic equivalent, syenite, are dominant rock types. Magmas of trachytic composition may be derived by fractional crystallization of basaltic magma, and Bailey and Schairer (1966) discuss various mechanisms whereby trachyte may subsequently differentiate on an oversaturated trend (to rhyolite) or an undersaturated trend (to phonolite).

The origin of trachytic magmas is discussed in Fitton and Upton (1987).

Other types

Arsoite—a leucotrachyte related to rhomb porphyry and kenyte; Arso-type trachyte.

Calcitetrachyte—contains > 10% calcite, probably primary.

Drachenfels trachyte (drakonite)—phenocrysts of sanidine and plagioclase in a groundmass of alkali feldspar microlites with accessory minerals (Drachenfels, W Germany).

Gibelite—abundant large phenocrysts of Na-microcline in a groundmass composed essentially of Na-microcline and showing flow texture.

Orthotrachyte—a silica-saturated trachyte.

Ponzite (ponzaite)—contains augite that may be rimmed with acmite or acmite-augite (Ponza Is., Italy).

Selagite—abundant phenocrysts of biotite in a holocrystalline groundmass mainly composed of orthoclase and diopside, but possibly quartz and olivine.

Shackanite—a analcime trachyte with rhombohedral phenocrysts of feldspar; also a variety of analcime rhomb-porphyry (Shackan, British Columbia, Canada).

Taimyrite—a nosean-bearing trachyte (Taimyr River, Siberia, U.S.S.R.).

Tutvetite—a trachytoid rock mainly composed of albite, microcline and much altered acmite and accessory pyrite (Tutvet, Norway).

IAN PARSONS

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Cross-references: *Intermediate igneous rocks; Phonolite; Rift valley volcanism; Syenite; Trachyandesite, trachybasalt; Volcanic rocks.*

TREND SURFACE ANALYSIS IN PETROLOGY

Trend surface analysis involves the recognition and measurement of trends in the properties of areally distributed samples. It is a type of regression analysis in which a relationship is sought between the values of a dependent variable (the property of interest) and two independent variables (geographical parameters). The type of data in which trends can be recognized and measured include the chemical or modal composition of rocks (e.g., % SiO₂, % hornblende), the composition of minerals, the specific gravity, or any other measurable property.

A trend surface is a plane or curved surface, defined by a mathematical equation that relates predicted values of the dependent variable to geographical parameters. The type of surface most commonly calculated is a polynomial surface. For example, if x and y represent the geographical parameters (e.g., E-W and N-S coordinates) and z represents the parameter whose variation is being studied (the dependent variable), the expression

$$z = a + bx + cy$$

corresponds to a first-degree polynomial surface, which is a flat plane. A second-degree polynomial surface would be of the form

$$z = a + bx + cy + dx^2 + exy + fy^2$$

i.e., it would contain terms of the second degree in x and y . This equation is of a curved (paraboloidal) surface. Higher-degree surfaces are progressively more convoluted and can be used to represent more complicated types of variation. Surfaces of 2nd, 3rd, 4th, . . . degree may be described as quadratic, cubic, quartic, . . . surfaces.

The deviation of measured values of the property from the values predicted by a calculated surface are called the residuals, and a measure of the variation of data about a surface is the sum of the squares of the residuals. In finding a surface of a given degree to represent a particular set of data, a least-squares method of calculation is used, i.e., the "sum of squares" is minimized. If z_{OBS} is an observed value of the dependent variable, z_{CALC} is the value predicted by a calculated surface at the same geographical location, and z_{MEAN} is the mean of all the observed values, then a measure of the "goodness of fit" of a surface is given by the expression

$$\frac{\sum (z_{OBS} - z_{MEAN})^2 - \sum (z_{OBS} - z_{CALC})^2}{\sum (z_{OBS} - z_{MEAN})^2} \times 100\%.$$

This is referred to as the percentage of the sum of squares explained by the surface. Percentage explanations as low as 10% or less may be significant if there is considerable local variation and the number of samples is large, although in many examples the percentage explanation is much higher.

Numerous computer programs have been written to carry out the lengthy calculations necessary in trend surface analysis, and several of these have been published. The technique is usually applied in a stepwise manner, in which surfaces of successively higher degree are fitted to the measured data and examined one by one to see whether they represent the variation any better than surfaces of lower degree. To find out whether a significant trend actually exists it is necessary to show that a trend surface fits the data better than a horizontal plane. To find out which degree of trend surface represents the trend best, it is necessary to find one which fits the data significantly better than surfaces of lower degree, and testing the significance of surfaces is usually done by analysis of variance (Harbaugh and Merriam, 1968). With irregularly spaced data, or data of which the residuals are not ideally distributed about the surface, it may be preferable to use the method of interpenetrating subsamples described by Agterberg (1964).

The purpose of calculating a trend surface may be either to recognize a regional trend, which is done by examining the trend surfaces, or to identify local deviations from a trend, which is done by examining the residuals.

Specific examples in which the technique has been used to study variations within a single igneous intrusion include (1) measurement of chemical and modal variations and heavy mineral content of a granite, (2) specific gravity variations in a peridotite as a measure of the degree of serpentinization, (3) variation in grain size in a diorite, and (4) variation in En-content of pyroxenes in an ultrabasic pluton. On a larger scale, Reed et al. (1983) used trend surface analysis to infer the polarity of a palaeo-subduction zone from compositional variations in a Mesozoic batholith. In most of these examples the investigators have used trend surface analysis to define and measure trends, but the examination of residuals has sometimes given additional information. For example, Whitten (1962) found that the deviations from the general trend of plagioclase/K-feldspar ratio in a granodiorite could be related to the "ghost stratigraphy" previously established by the mapping of xenoliths, even though no contamination was obvious in most of the actual samples analysed.

The trend surfaces calculated in the examples

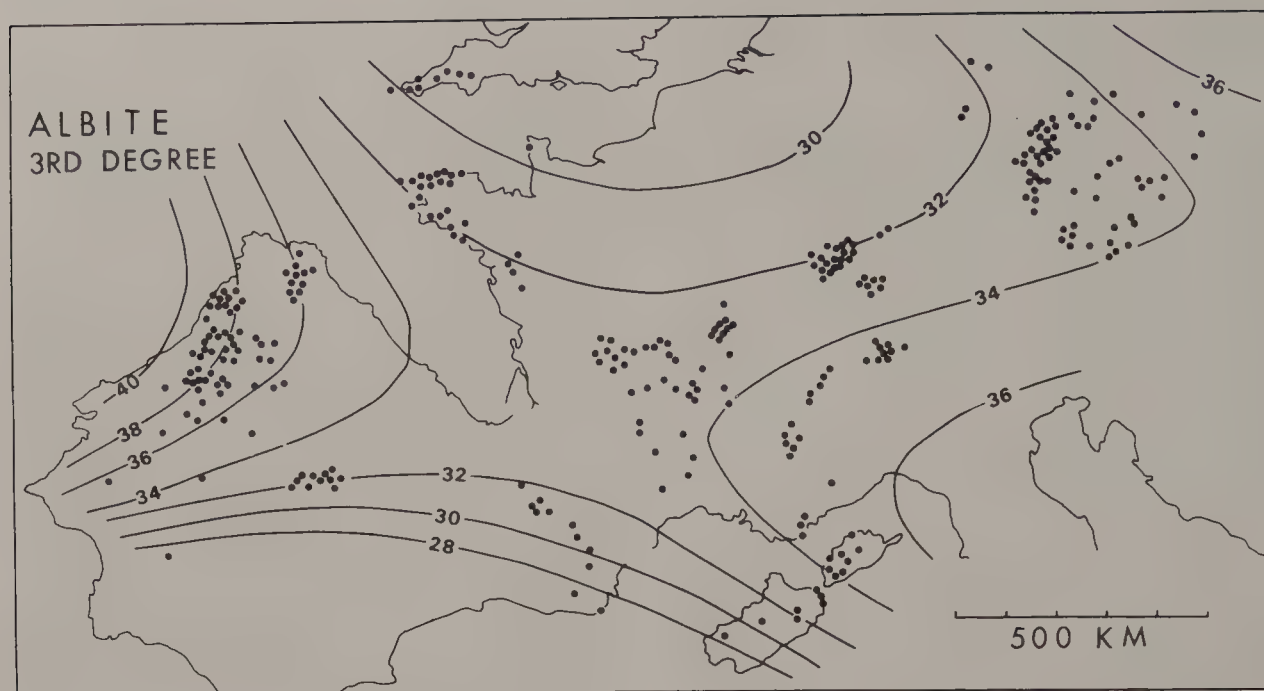


FIGURE 1. Trend surface of third degree for normative Ab as a percentage of $Q + Or + Ab$ in the Variscan (Hercynian) granites of W Europe; the surface distinguishes the main trend of regional variation from considerable local variations in composition, and is used to relate the conditions of magma generation to regional variation in the geothermal gradient. (After Hall, 1973a.)

quoted above have been used to throw light on such processes as differentiation, contamination, or metasomatism within an intrusion. Trend surfaces have also been used to measure the bulk composition of rock masses, by integrating the volume under a surface, and to measure variation between intrusions, as for example the granites of the Caledonian and Variscan orogenic belts (Fig. 1). Extension of trend analysis to variation in three dimensions permits the examination of vertical as well as horizontal variations within an intrusion—the term hypersurface has been applied to three-dimensional contour surfaces.

The application of trend surface analysis to the distribution of trace elements, especially those of economic interest, is frequently hindered by their extremely asymmetrical frequency distribution of concentrations, and in such cases a logarithmic transformation of the data has been found useful. A more general distribution-free approach is the calculation of median surfaces, representing the variation in the median value of the dependent variable rather than its mean value (Hall, 1973b). Agterberg (1969) and Miesch and Connor (1968) have described the use of techniques other than polynomial regression to represent a real variation.

A review of trend surface analysis, with numerous examples, is given by Harbaugh and Merriam (1968).

A. HALL

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Cross-references: *Petrochemical calculations; Statistics in petrology.*

TUFFISITE

The country rocks at or near to the margins of many volcanic necks (pipes) or other channels of eruption are penetrated by intrusive fine-



FIGURE 1. Brecciated Jurassic marl penetrated by basaltic intrusive tuff ("tuffisite" as originally defined); Swabia, W Germany. (After Cloos, 1941.)

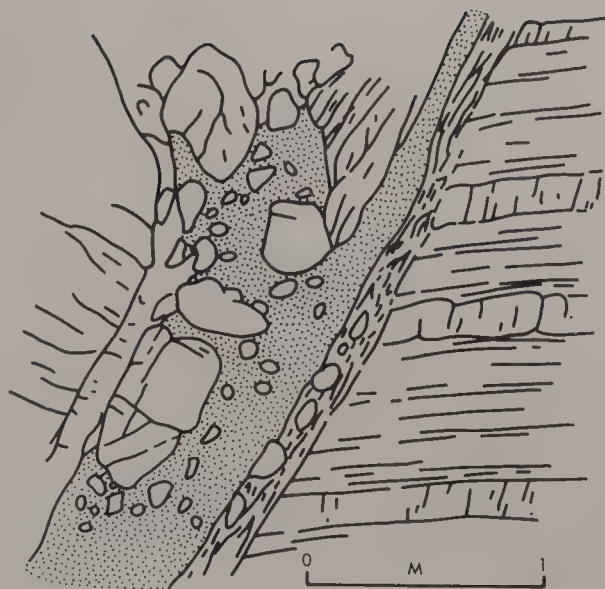


FIGURE 3. Tuffisite dyke in a fault in Jurassic limestones; Swabia, W Germany. (After Cloos, 1941.)

grained pyroclastic material. This consists generally of mixtures of fragments derived in various proportions both from the country rocks themselves and from the parent magma. The penetration may take the form of ramifying matrices in brecciated country rock (Fig. 1) or of discrete intrusions emplaced along joint, fault, or bedding planes (Fig. 2). There are records of gradational passages between such breccias and the pyroclastic rocks filling the central parts of necks: similarly, both have been described as passing without break into discrete clastic intrusions. Of such intrusions, dykes (Fig. 3) tend to be more common and to extend farther from source than either sills or irregular apophyses. Emplacement under pressure is frequently indicated by the flow-alignment of the long axes of particles parallel and adjacent to the containing walls or to the surfaces of larger contained xenoliths.

Cloos (1941) originally introduced the term

"tuffisite" for the tuff-penetrated breccias, quoting them as evidence of a process whereby pipes are drilled by gas-tephra streams that advance ahead of ascending columns of magma. He envisaged that from the country rocks, which are first fractured and penetrated by the mobile gas and tephra, blocks are detached by stopping into the streams, where they become progressively comminuted and absorbed by admixture with magma fragments. Reynolds (1954) compared the process ("tuffisitization") with fluidization processes in the chemical engineering industry, at the same time broadening the meaning of tuffisite to include dykes, sills, and apophyses of intrusive tuff. It is in this sense, as a synonym for intrusive tuff, that the term tuffisite has become most widely accepted and used; in consequence, the rock type originally so designated by Cloos has become "tuffisitic breccia" (Francis, 1977).

E. H. FRANCIS

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Cross-references: *Breccia pipe*; *Cryptovolcanic structure*; *Diatreme*; *Fluidization*; *Volcanic neck*.



FIGURE 2. Borehole core of Carboniferous sandstone (pale) penetrated by intrusive tuff (dark); Fife, Scotland.

U

ULTRABASIC IGNEOUS ROCKS

Ultrabasic is a general term, probably first used by J. W. Judd in 1881, applied to igneous rocks containing little or no feldspar and characterized essentially by the abundance of one or more common mafic minerals such as olivine, pyroxene, and amphibole. Chemically, the rocks contain $<45\%$ SiO_2 and the term ultrabasic is used along with basic, intermediate, and acid(ic) in a chemical classification of igneous rocks as distinct from ultramafic, mafic, and felsic which refer to a mineralogical color index. However, the term ultrabasic commonly is used incorrectly as synonymous with ultramafic to include such rocks as ultramafic pyroxenites that contain as much as 55% SiO_2 .

Daly (1933) recognized three clans of ultrabasic rocks—(1) gabbroic and their extrusive equivalents, (2) feldspathoidal, and (3) ultramafic. For the first clan, the subdivision is not a natural one because it cuts across other subdivisions that have petrogenetic significance. It includes basaltic members of both the hypersthene-normative tholeiitic suite and the nepheline-normative alkaline suite (Fig. 1) but with more ultrabasic types in the former than the latter (Irvine and Baragar, 1971). This means that some but not all *ankaramites*, *oceanites*, *boninites*, and *picrites* are considered to be ultrabasic. Similarly, only the most calcic *anorthosites* are ultrabasic, although, like the less calcic varieties, they are felsic on a color

index. Gabbro does not qualify for this clan, but a calcic *troctolite* might.

The feldspathoidal clan should strictly include *carbonatites*, but is usually restricted to more siliceous peralkaline rocks. These would include rocks as diverse as melilitites, some *lamprophyres*, *limburgites*, *ijolites* and *nephelinites*. The ultramafic clan is the least satisfactory and most confusing of this grouping because it is not synonymous with the ultramafic color index category. It includes most olivine-dominated *peridotites* (such as *dunite*), *glimmerites*, *hornblendites*, *komatiites*, *kimberlites*, and *meimechites* (or *meimechites*; extrusive equivalents of kimberlite), but not pyroxenites unless they carry sufficient ore minerals in the mode for the $\%$ SiO_2 in the chemical analysis to be <45 .

As the term ultrabasic has little petrogenetic significance, ultrabasic rocks can occur virtually anywhere basic or ultrabasic alkaline igneous rocks are found. This is illustrated by the diversity of geologic environments in which volcanic members of the ultrabasic group occur: nepheline basalts, ankaramites, and oceanites among the present-day Hawaiian basalts erupted in the Pacific Ocean; alkali olivine basalts among the Cenozoic lavas of Japan in the circum-Pacific volcanic belt; nepheline-, leucite-, melilite-, and kalsilite-rich Cenozoic lavas in the rift valleys of E Africa; meimechites in the Triassic flood basalts of Siberia, U.S.S.R.; and komatiites from the Archaean greenstone belts of S Africa and other continents. Ultrabasic plutonic masses commonly result from processes of fractional crystallization and crystal accumulation, including those of both tholeiitic and alkaline character in E Greenland (Skaergaard and Lilloise intrusions, respectively). Others occur in zoned alkaline complexes such as the Monteregian plutons of the Montreal district, Canada, and the many ultrabasic alkaline rocks in the U.S.S.R. recorded by Vorobieva (1960). In addition, the abundant serpentinites of ophiolite assemblages are generally ultrabasic as well as ultramafic. These and other representatives of upper mantle material generally have SiO_2 less than or ca. 45% (see **Pyrolite**).

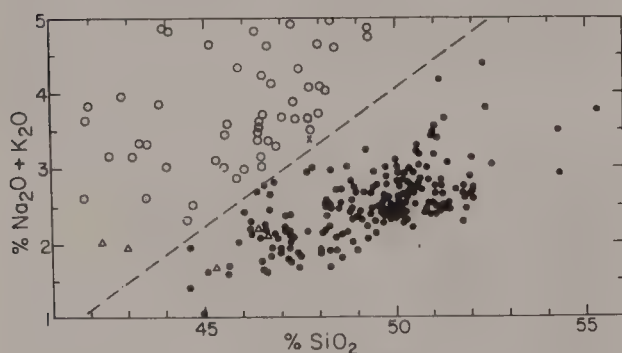


FIGURE 1. $\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. SiO_2 plot showing relative abundances of ultramafic rocks in the tholeiitic suite (solid circles) and alkaline suite (open circles), Hawaii. (From Macdonald and Katsura, 1964.)

D. R. BOWES

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- Cross-references: *Alkaline rocks—undersaturated; Anorthosite; Basic igneous rocks; Gabbro and norite; Igneous rocks—classification and nomenclature; Kimberlite; Lamproite; Lamprophyre; Limburgite; Magmatic differentiation; Melilitite; Mid-ocean ridge and ocean-floor petrology; Ophiolite; Peridotite; Picrite; Plutonic rocks—classification and nomenclature; Pyrolite; Pyroxenite; Ultramafic lavas; Ultramafic rocks.*

ULTRAMAFIC BELTS

A characteristic feature of many orogenic zones is the presence of ultramafic rocks as lenses, sheets, or irregular bodies parallel to the structural trends of the orogen and commonly extending discontinuously for upwards of 1000 km along the strike of the orogen. Such a linear or en-echelon series of ultramafic bodies is called an *ultramafic belt*, or, more loosely, a “serpentine belt” or “ultrabasic belt.” The rock types present in such bodies are of igneous or metamorphic character and include peridotite, dunite, pyroxenites, serpentinites, and gabbroic rocks. A common but not invariable field association is with layered and massive gabbro, and dolerite dykes as a parallel-sheeted complex, overlain by pillow lavas of spilite and altered basalt, graywacke, and chert; the total lithological association is referred to as the “ophiolite suite” (Coleman, 1977). In contradistinction to the petrographically similar rock types occurring as differentiates of basaltic magma in stable, nonorogenic regions, the rocks of the ultramafic belts are referred to as peridotites, etc., of “*alpine-type*” or of “*orogenic type*.”

Structural relations

Hess (1955, and earlier papers) considered that ultramafic belts typically consist of two parallel series of bodies about 190 km apart located along opposite flanks of a central zone of most intense orogenic deformation (Fig. 1). He noted a parallelism of ultramafic belts with the negative gravity anomaly strip of island arcs and suggested that ultramafic rocks were emplaced at the first great deformation of an orogen but did not accompany later deformations and episodes of intrusion.

Structurally, serpentinitized peridotites not uncommonly form the cores to anticlines but may also form sill-like bodies on the limbs of folds. Bedding or foliation in country rocks is commonly parallel to the margin of ultramafic bodies, although discordance or deflection of country rock trends may be found locally. Typical contacts between ultramafic and country rock are strictly tectonic rather than igneous, showing mylonitic texture in fresh ultramafic rocks, shear zones in serpentinite, thin breccia zones of serpentinite and country rock or blocky, fractured serpentinite with common slickenside surfaces. Many alpine ultramafic rocks occur along a major structural or lithologic break, e.g., (a) a major fault of transcurrent or overthrust type, (b) between an older metamorphic terrane and lower-grade or nonmetamorphic terrane, (c) between a metamorphic “sialic” block and a block of eugeosynclinal and basic igneous character, or (d) as a flat or gently dipping thrust plate.

In modern theories of “continental drift,” “sea-floor spreading,” or “plate tectonics,” the ultramafic belts, particularly those with structural setting of (c) and (d) have been attributed to continental crust–oceanic crust collision with elimination of oceanic crust by subduction beneath the continental crust or, more rarely, by overthrusting of oceanic crust (including basic and ultramafic rocks, etc.—the ophiolite suite) on to a sialic continental or island arc crust. Ultramafic belts have been interpreted as marking suture lines showing the former sites of subduction of oceanic crust and lithosphere.

On the Archaean shields of Australia, Canada, S Africa and Zimbabwe, greenstone belts that consist largely of volcanic products (particularly andesite and basalt), with some admixed sediment, also contain ultramafic rocks. Some of these are of extrusive character, preserving volcanic structures and igneous quench textures, and having compositions ranging from picritic to peridotitic (Table 1; Viljoen and Viljoen, 1969). They provide unequivocal evidence for the extrusion of ultramafic liquids (peridotitic komatiite) at



FIGURE 1. Ultramafic belts of western N America.

temperatures $>1600^{\circ}\text{C}$ at a very early period in the Earth's history, but such peridotitic magmas are apparently not present among Palaeozoic, Mesozoic, or Cenozoic ultramafic belts.

Primary mineralogy

The ultramafic rocks are rich in magnesian olivine (Fo_{88-95}), the most typical rock type being peridotite with 60–80% olivine, 20–40% enstatite and accessory red-brown chromite and commonly bright green chrome-diopside. Chem-

ically, the ultramafic rocks have very high MgO contents, are very low in Na_2O , K_2O , Al_2O_3 , and usually low in CaO and SiO_2 (Table 1). Rocks composed entirely of enstatite or bronzite may occur, but clinopyroxene-rich rocks are normally of minor importance. Chromite occurs disseminated but also as schlieren, pods, and irregular bodies approaching monomineralic composition. Ultramafic rocks may pass abruptly or transitionally into troctolites and gabbros, also typically with high Mg/Mg + Fe atomic ratios. The proportion of

peridotites under conditions of very high pressure (see **Upper mantle—experimental petrology**). The olivine + enstatite + diopside + spinel assemblage is found in peridotites having high-temperature aureoles (hypersthene + augite granulites), as at Tinaquillo (Venezuela), Lizard (Cornwall, England), Mt. Albert (Quebec, Canada), Beni Bousera (Morocco), Serrania del Ronda (Spain), and probably Lherz (Pyrenees). In such rocks the spinel is rich in the MgAl_2O_4 end-member (40–60 wt. % Al_2O_3) and both enstatite and diopside are rich in Al_2O_3 . This mineral assemblage is also characteristic of the peridotite xenoliths occurring in alkali basalts and is representative of the mineralogy of the upper part of the mantle lithosphere.

The olivine + enstatite + diopside + garnet assemblage is found in lenticular peridotite bodies in some deeply dissected orogenic belts such as Norway, Germany, Czechoslovakia, Italy, Switzerland, and Austria. The garnet of this assemblage is very rich in the pyrope molecule (60–75% pyrope), the enstatite is low or very low in Al_2O_3 and CaO content, and the clinopyroxene is a diopside with appreciable Cr_2O_3 and jadeite content. This mineral assemblage is characteristic of xenoliths of the diamond pipes of Africa and the U.S.S.R. and is representative of a deeper zone within the mantle lithosphere than the spinel-bearing peridotites.

Mineralogical banding is present in almost all peridotites. In a few cases such as eastern Papua, New Zealand, SE Alaska, such banding is demonstrably due to crystal settling and accumulation in a highly fluid magma. In such cases rhythmic repetition of mineralogically varying layers, poikilitic textures, and euhedral crystals point to the igneous origin of the banding. However the most typical compositional banding has the character of flow banding developed by recrystallization and flow of initially inhomogeneous material in the solid or near-solid state. The initial cause of the inhomogeneity can seldom be specified. The rocks have granoblastic or lepidoblastic textures, or more rarely and particularly in the olivine-rich rocks, have mylonitic texture.

Secondary mineralogy

Peridotites and dunites are usually altered, partly or completely, to serpentinite (Table 1, analyses 2, 4). Serpentine minerals have the ideal formula $\text{Mg}_3\text{Si}_2\text{O}_5(\text{OH})_4$ with variable substitution of Fe, Ni and Al, but structurally may be one of several varieties of chrysotile (orthochrysotile, clinochrysotile, parachrysotile), or lizardite, or antigorite. Ser-

pentinities consisting of chrysotile–lizardite mixtures commonly show a texture in which olivine is pseudomorphed by serpentine with mesh-texture (“*fenster struktur*”) and orthopyroxene is pseudomorphed by plates of “bastite serpentine.” In contrast, serpentinites consisting of antigorite commonly have a completely recrystallized texture with random orientation of flakes and laths of antigorite replacing preexisting olivine and enstatite or mesh-texture + bastite serpentinites.

Serpentinites have a density of 2.6–2.8 g/cm³ in contrast to that of fresh dunite or peridotite of about 3.3 g/cm³. Serpentinization of olivine by addition of H_2O with SiO_2 or CO_2 without removal of MgO would involve a volume increase of about 60%. In contrast, serpentinization at constant volume requires removal of very large amounts of MgO in solution. The occurrence of brucite in serpentinites and the retention of preserpentinization Mg/Si, Mg/Ca, Mg/Al, etc. ratios (deduced from primary mineral compositions and preserpentinization modal mineralogy) has demonstrated that serpentinization occurs by addition of water at $T < 500^\circ\text{C}$, with little change in Mg, Si, Fe, Al, Cr, etc., but commonly with depletion in more mobile elements such as Na, K, Ca, and associated trace elements.

Various hypotheses have been put forward to account for the source of water for serpentinization.

1. Hess proposed the existence of a *low-temperature magma of serpentinite composition* (i.e., containing ca. 13 wt. % H_2O) that would first crystallize olivine and pyroxene; on further cooling, the highly water-enriched residual liquid would convert these minerals to serpentine. This hypothesis is inconsistent with experimental work of Bowen and Tuttle (1949) on the system $\text{MgO-SiO}_2\text{-H}_2\text{O}$ which demonstrated the lack of a liquid phase and the stability of olivine + enstatite crystals with water vapor at temperatures up to $1,000^\circ\text{C}$.
2. *Later granitic intrusions* with high water content could provide water for serpentinization of preceding peridotite bodies. This fails as a general explanation but may be of local significance, particularly in producing bodies of talc, tremolite–actinolite and anthophyllite from peridotites.
3. The *country rock* provided the water. This is particularly applicable to serpentinite bodies in unmetamorphosed or low-grade metamorphic terranes and is probably the most common source of water.
4. *Water vapor* moving along fault planes and lithologic discontinuities was the source.

Associated rocks

Bodies of talc, anthophyllite, and magnesite are derived by metasomatic action on the ultramafic rocks. In addition, there are often bodies of *rodingite* consisting of grossular or hydrogrossular with vesuvianite, chlorite, diopside, or prehnite. These rocks are derived by metasomatic action on small bodies of gabbro or, less commonly, of country rock. Small bodies of albitite or quartz-albite rocks may have a complementary relation to rodingite and both are probably formed during the period of serpentinization.

Economic significance

1. Very pure and Mg-rich dunites are quarried in Norway and U.S.A. for use as refractory and casting sands, and for manufacture of olivine (Fo_{90-95}) or pure forsterite (Fo_{100}) refractory bricks.
2. Economic deposits of talc and of magnesite are locally derived from alpine ultramafic rocks.
3. Asbestos fiber (usually of chrysotile mineralogy but in some cases of anthophyllite) is mined from ultramafic bodies, usually those of greater age and more complex history.
4. Major chromite deposits occur in alpine ultramafic masses in Turkey, the Balkan Peninsula, Phillipines, and Cuba.
5. Where ultramafic rocks have been subjected to deep weathering under conditions of lateritization, Ni is leached out and deposited in the lower zones of the soil profile, commonly as the hydrated silicate, garnierite. Such Ni-enriched zones of laterite provide deposits of major economic significance in Cuba and New Caledonia. Nickel sulfide deposits occasionally also occur in association with ultramafic rocks of alpine type but are typically associated with the Archaean ultramafic magmas of the shield regions of Australia, Canada, and Zimbabwe.
6. Pt, Pd, Os, Ir, Rh, and Ru occur alloyed in varying proportions and in deposits of varying significance in placer and alluvial deposits derived from ultramafic terranes in the Urals, western N America, Colombia, Australia and other areas.

Origin and emplacement

A hypothesis of derivation of the alpine ultramafic suite by differentiation at the level of emplacement from a basaltic magma in the manner of the Bushveld, Stillwater, and Rhum intrusions, is not supported by field, petrographic, or chemical evidence.

Geophysical and geochemical data supports the concept that the Earth's mantle, below the Mohorovičić discontinuity, is of peridotitic composition (see **Pyrolite**). Thus, ultramafic rocks are present beneath continental areas at depths of 25–45 km. In areas of oceanic crust, peridotite (with a seismic velocity $V_p = 8$ km/sec.) is present 5–6 km below the ocean floor, and is overlain by the oceanic crust consisting of a layer 4–5 km thick with $V_p = 6.4\text{--}6.9$ km/sec. corresponding probably to basaltic rocks and some associated serpentinized peridotite and a thin layer, usually <1 km thick, of sedimentary material.

The nature of peridotite contacts, the strain effects, the flow banding, the variety of mineral assemblages (reflecting in some cases very high pressure at crystallization), and the structural position of many peridotites, all stress the essential role of solid-state deformation in emplacement of the bodies. The data at present force the conclusion that a single mechanism of emplacement is inadequate for all alpine ultramafic bodies, and emplacement of individual bodies may be in part or completely by one of the following mechanisms.

1. Tectonic or structural exposure of sub-oceanic mantle material by major faulting or folding: this mechanism is particularly important in the context of subduction of oceanic crust and lithosphere in regions of active orogenesis: it is particularly applicable to the very large peridotite-gabbro complexes (the ophiolite complexes) such as in Cuba, Puerto Rico, Phillipines, Borneo, Celebes, Papua-New Guinea, Solomon Islands, New Hebrides, New Caledonia, and Cyprus. These areas may preserve rocks of igneous aspect or of metamorphic aspect and commonly show evidence of multiple phases of intrusion and deformation.

2. "Cold intrusion" or re-intrusion of lenticular serpentinite bodies by movement on sheared margins and multiple slickenside surfaces within the serpentinite: this type of serpentinite is commonly found in low-grade metamorphic or nonmetamorphic rocks.

3. Intrusion or re-intrusion of lenticular un-serpentinized dunites and peridotites by internal deformation and recrystallization at temperatures similar to those of the country rock: this type of peridotite occurs in high- or medium-grade metamorphic terranes and shows no evidence of a thermal aureole but a high degree of recrystallization, crystal orientation, and flow banding or foliation. Such bodies are occasionally sheathed by sheared serpentinite or by fault contacts and in part may owe their emplacement to processes 1 and 2. Peridotites of this type preserving internal bands of garnet

peridotite or lenses of eclogite imply "cold" movement from deep crustal levels or imply uplift and erosion to expose metamorphic terranes formed by deep crustal subduction.

4. Emplacement of high-temperature but crystalline diapirs such as Lizard, Cornwall, England, Tinaquillo, Venezuela, Mt. Albert, Quebec, Canada, Beni Bousera, Morocco, and Serrania del Ronda, Spain. Such peridotites probably began movement in the mantle as partially molten peridotite diapirs but final emplacement at crustal levels was as material of crystalline or possibly crystals + interstitial fluid character. Peridotites of this character have a distinct dynamothermal aureole with development of rocks of granulite facies adjacent to the peridotite.

5. Emplacement of liquid or partially liquid magmas of mafic to ultramafic character along the same channel giving zoned basic-ultramafic complexes such as those of SE Alaska.

6. Extrusion or intrusion of extremely high-temperature peridotite magmas (peridotitic komatiites) at $>1,600^{\circ}\text{C}$ during Archaean volcanism. This phenomenon is apparently restricted to very early stages of the evolution of the Earth and may imply very different geothermal gradients in the primitive Earth or more catastrophic and rapid convective motions in the Earth's upper mantle.

The factor common to these mechanisms seems to be the necessity for the large directed stress present in orogenic zones to bring ultramafic material of relatively high density normally present in the mantle to emplacement and exposure at crustal levels. The presence of ultramafic rocks as such a constant component of orogenic belts is evidence of the participation of both mantle and crust in the orogenic process.

D. H. GREEN

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Cross-references: *Basic igneous rocks*; *Greenstone*; *Mafic and ultramafic inclusions in igneous rocks*; *Ophiolite*; *Peridotite*; *Serpentinite*; *Ultramafic lavas*; *Ultramafic rocks*; *Upper mantle—experimental petrology*.

ULTRAMAFIC LAVAS

Whether ultramafic rocks crystallized from ultramafic liquids provided the subject for a controversy that raged for nearly half a century. Bowen (1928) considered that all ultramafic rocks were the products of the fractional crystallization of basic magma, and that the temperatures necessary ($>1,400^{\circ}\text{C}$) precluded the existence of ultramafic magmas in the crust (Hess, 1955). Nevertheless, throughout this period various workers cited instances of ultramafic material occurring as dykes and sills, or even lavas (cf., Bailey and McCallien, 1953; Searle and Vokes, 1969). These were interpreted as basalts enriched in olivine phenocrysts or mobilized cumulates. In 1969 the first evidence for peridotitic liquids having been erupted at the Earth's surface was published by Viljoen and Viljoen, who had studied the Archaean rocks in the valley of the Komati River in the Barberton Mountain Land greenstone belt, Transvaal, S Africa. Peridotitic pillow lavas were recorded as being interlayered with a basaltic extrusive sequence and sills (Fig. 1). The ultramafic lavas were called komatiites and subsequently have been recognized in Archaean greenstone belts all over the world. Isolated examples of these and other ultramafic lava types have also been documented in Proterozoic and Phanerozoic rocks, usually ophiolites, or in ocean-ocean island arcs (see review papers in Arndt and Nisbet, 1982).

Komatiite

Komatiite is an ultramafic rock erupted as lava and forming both pillows and massive flow units. As lavas, their characteristic features include chilled flow tops, polyhedral jointing, well-developed spinifex texture, pillows, fragmental structures (tuffs and breccias), and abundant glass. The ubiquitous presence of glass precludes an unequivocal normative mineralogy being defined, but features indicating an ultramafic parental liquid are a

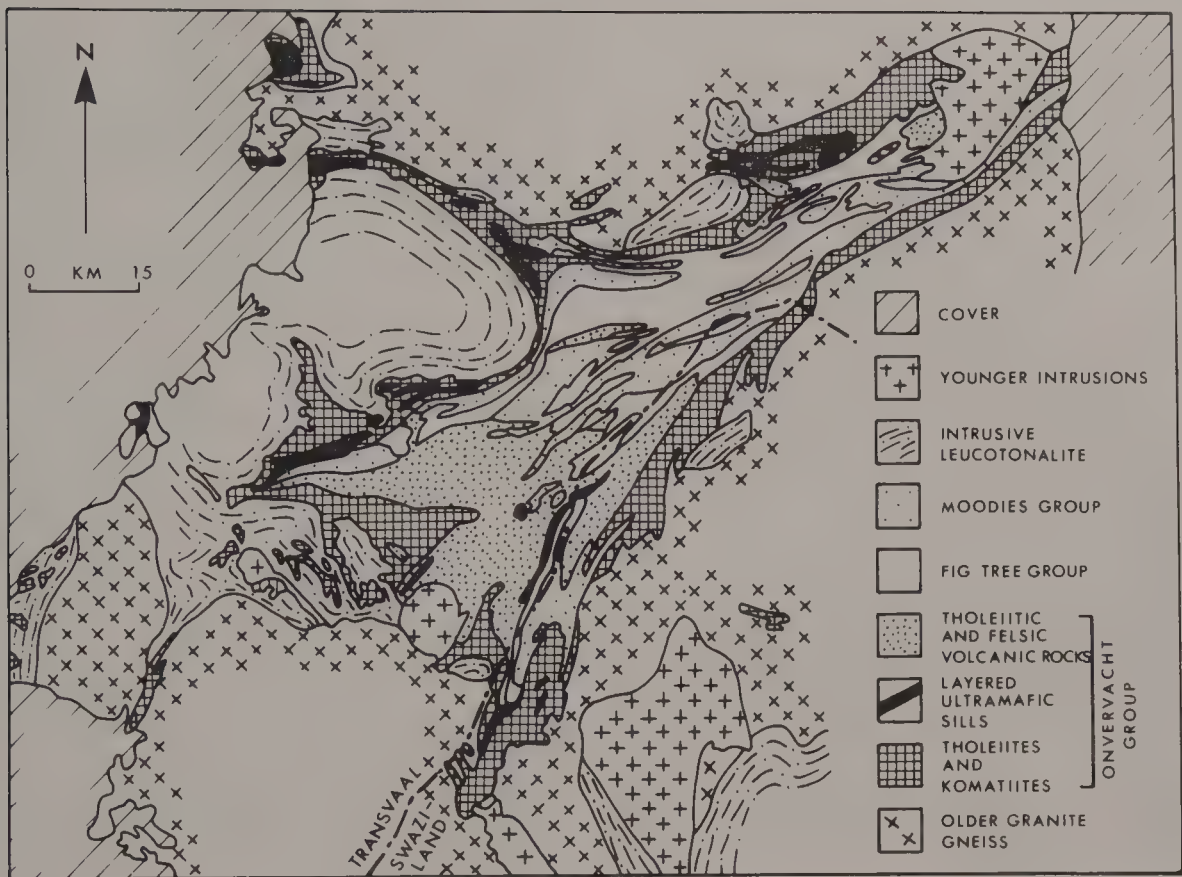


FIGURE 1. Distribution of komatiitic lavas in the Archaean greenstone belt, Barberton Mountain Land, Transvaal, S Africa.

predominance of olivine and pyroxene phenocrysts and their high Mg content, indicating equilibrium with a highly magnesian liquid at high temperatures (usually $>1,400^{\circ}\text{C}$).

Fresh komatiite is a contradiction in terms. The once glassy matrix is invariably heavily altered to talc, serpentine, and chlorite, while phenocrysts are usually pseudomorphs of amphibole, talc, or serpentine. Chemical criteria have therefore to be applied with caution. Chemical characteristics that may be diagnostic include higher Mg, Ni, and Cr contents and higher $\text{CaO}:\text{Al}_2\text{O}_3$ ratio than for otherwise comparable picritic basalts and ankaramites, and lower TiO_2 content than for such rocks with a comparable SiO_2 content. When the whole-rock composition is recalculated as being anhydrous, the MgO, FeO, CaO, and SiO_2 contents are equivalent to 20–35% modal olivine or 15–20% normative olivine, and 65–70% normative mafic minerals.

Komatiitic basalts

These rocks have chemical affinities with komatiites but petrologic affinities with tholeiitic basalts. They have a spatial association with komatiites, and within such volcanic sequences

there is often a progressive change from basaltic to ultramafic compositions. Spinifex textures (both olivine and pyroxene) are common and the type may relate to composition; olivine spinifex occurs in komatiitic basalts with MgO between 12 and 28 wt.% and pyroxene spinifex occurs in the range 9–18 wt.% MgO. The presence of highly magnesian pigeonite and unusually sodic (for basalt) plagioclase may be diagnostic. Compared to picritic basalts and oceanites these rocks have lower Ti, Nb, Y, Zr, and light REE (rare-earth element) abundances, and higher Mg, Ni, and Cr. When compared to basalts of similar SiO_2 content they are seen to have lower alkali content, higher $\text{CaO}:\text{Al}_2\text{O}_3$ ratio, and higher MgO content. However, there is considerable overlap in composition and mineralogical properties between these rocks and the high-Mg melabasalts and picrites associated with other basaltic provinces. If it were not for the spatial association with komatiites in greenstone belts it is unlikely they would have been separated as a distinct group. Many workers in greenstone terrane have used the term “komatiite” to include all the mafic and ultramafic lavas with spinifex textures, but the classification used in Canada (Fig. 2) is now widely accepted and

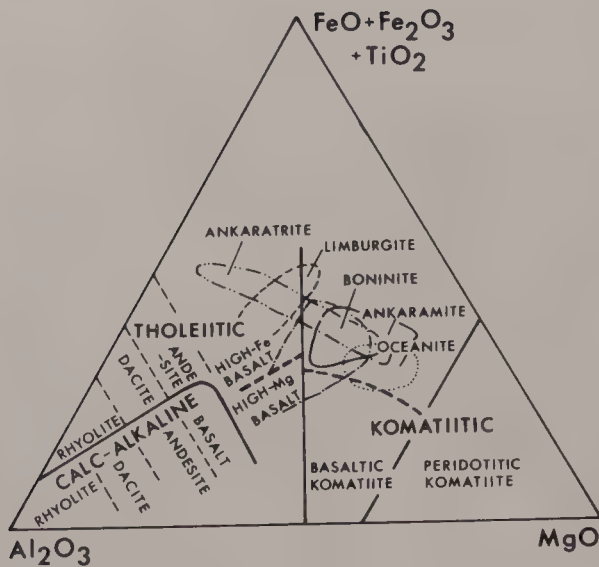


FIGURE 2. Jensen's cation percentage plot for distinguishing komatiites from basalts (after Arndt and Nisbet, 1982); data for boninite from Cameron and Nisbet (1982, in Arndt and Nisbet, 1982); data for ankaramite, ankaratrite, limburgite, and oceanite from Sørensen (1974).

separates komatiitic basalts from the ultramafic or peridotitic komatiites.

Other ultramafic lavas

Boninite is an olivine-clinoenstatite or bronzite-clinopyroxene phyrlic rock with a glassy matrix. The pyroxenes form acicular or elongate prismatic plates, and the olivines are generally stubby, subhedral prisms, although spinifex textures are also known. SiO_2 content is higher than basalt (56–60 wt.%), but MgO is also high (10–15 wt.%). For their SiO_2 content they have low TiO_2 and alkali content and high total Fe content and $\text{CaO}:\text{Al}_2\text{O}_3$ ratio. As a result, normative feldspar is less than 10%, and never appears in the mode. They are high-Mg andesites. *Limburgite* is a glassy olivine-clinopyroxene phyrlic lava with high alkali and relatively low Al_2O_3 content. The norm includes nepheline, though rare feldspar phenocrysts are known. *Ankaratrite* is a biotite-bearing olivine melanephelinite. *Ankaramite* is an augite metabasalt. *Oceanite* is an olivine melabasalt in which there is more than 50% modal olivine. *Raqqaite* has the composition of a pyroxenite. *Meymechite* (meimechite) is an ultramafic lava consisting of 50–60% olivine phenocrysts in a matrix of glass with microlites of titanite (not plotted on Fig. 2 because of alkaline nature). Accessory minerals include perovskite, and the rock has affinities with kimberlite.

Occurrence

The secular occurrence of komatiites and komatiitic basalts is well constrained, but the

spatial occurrence, especially their tectonic setting, is not. Well over 80% of documented komatiites are in Archaean greenstone belts, with over 90% occurring in greenstone belts in the early Precambrian in general (early Proterozoic examples occur in S America and N Finland). They form a minor component in a basic igneous province in the early Proterozoic Ungava-Cape Smith fold belt of Quebec, Canada, and more recent occurrences of true komatiites are documented in the Ordovician ophiolite of Newfoundland. Rocks similar to komatiitic basalts are known from a number of Phanerozoic orogenic belts and ophiolite complexes including the Othris ophiolite, Greece, the Troodos ophiolite, Cyprus (both Mesozoic), the Cretaceous volcanic rocks of Wairapapa, North Island, New Zealand, and the Eocene ophiolite of Gorgona Island, Colombia (Fig. 3; see review by Cameron and Nisbet in Arndt and Nisbet, 1982).

Boninites, that form part of a suite that includes komatiites and komatiitic basalts, are rare but widespread members of the volcanic suite developed in the ocean-ocean island arcs of the western Pacific Ocean. Similar rocks are known from the Troodos ophiolite (Cyprus) and from Newfoundland.

Limburgites, ankaramites, ankaratrites, and oceanites occur as minor components of ocean island volcanic suites (e.g., Canary Islands and Reunion), continental flood basalt provinces (e.g., Karroo, S Africa) and rift valley volcanic provinces (African rift, Rhine graben, W Germany). Meymechites (meimechites), that

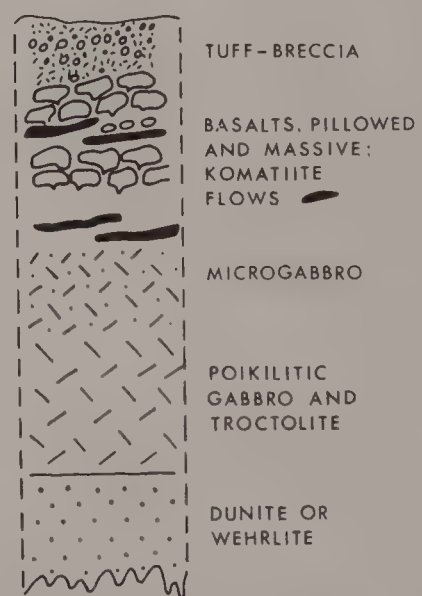


FIGURE 3. Schematic cross-section of the Eocene Gorgona Island ophiolite, Colombia, showing the position of the komatiite flows; after Echeverría in Arndt and Nisbet (1982).

may be extrusive kimberlites, are only known from the Triassic flood basalt plateau of the Siberian platform (Meymech River), U.S.S.R.

Komatiite lava flows

Many komatiites form pillow lavas and are associated with large amounts of autoclastic or hyaloclastic tuff, indicating eruption into shallow water. However, in W Australia, in the Barberton Mountain Land, S Africa (Fig. 1), and the Rouyn-Noranda and Munro areas of Quebec and Ontario, Canada, great thicknesses of massive komatiite flows interlayered with basalts have been recorded. The thickness of individual flow units may exceed 10–20 m with a semiregular internal zonation shown. This zonation, which varies with overall composition and flow thickness, and between localities, is expressed by cooling phenomena, spinifex mineralogy and morphology, overall petrography, and epiphenomena such as sulfide accumulation (Fig. 4).

The spinifex texture, and the composition of the phenocrysts involved indicate the eruption of a virtually crystal-free lava that subsequently chilled rapidly in its margins. However, in the more massive flows, total heat loss may have

been sufficiently impeded to permit turbulent convection to continue within the flow for some time after eruption. The lava itself may have had an exceptionally low viscosity, permitting individual flows to travel perhaps 50 km or more over very shallow gradients (Huppert et al., 1984). Where large amounts of sulfur were present, either indicating a high-S magma, or the assimilation of S-rich sediment, an immiscible, high-density sulfide liquid, rich in Cu, Fe, and Ni, may have separated as droplets that accumulated in topographic depressions at the base of the flow (Usselman et al., 1979). Thus, although rapid chilling produced the upper spinifex-textured and initially glass-rich zone in lava units, the slow-cooled and highly fluid flow interior may show considerable complexity. In some cases, crystal fractionation may have occurred, and some Canadian flows display a crude layering, including the presence of gabbroic layers (Hynes and Francis in Arndt and Nisbet, 1982).

Genesis of komatiites

The association and chemical similarities of komatiites, komatiitic basalts, high-Mg tholeiites and boninites suggest a petrogenetic

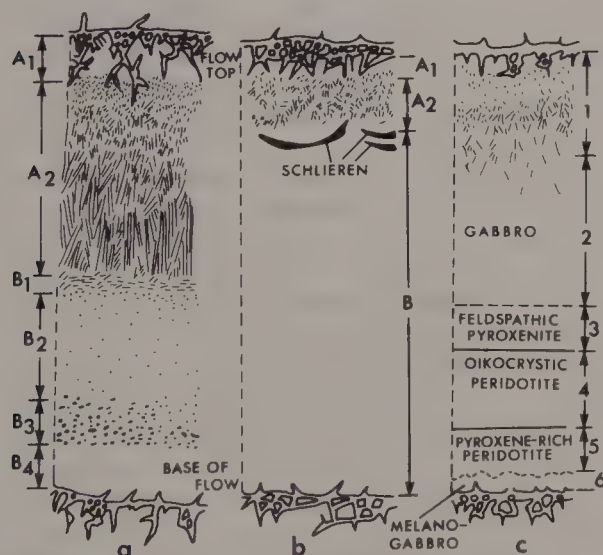


FIGURE 4. Layering in komatiite lava flows.

(a) Lava with thick spinifex-textured zone.

- A₁ Chilled, fractured and brecciated flow top, frequently with hyaloclastic debris; the lava itself consists of microspinifex olivine or pyroxene in glassy matrix.
- A₂ Massive spinifex zone with top of randomly-orientated skeletal grains merging downward into predominantly perpendicular, very coarse platy spinifex; the matrix is largely fine-grained, possibly devitrified glass.
- B₁ Foliated spinifex, skeletal, or platy phenocrysts with preferred (flow-oriented) direction parallel to flow top.

B_{2,3} Downward coarsening peridotite that may have cumulate textures.

B₄ Fine-grained peridotite with glass and microspinifex at the base; if the olivine or pyroxene in zones B_{2,3} and B₄ are elongate, they may show a weak foliation.

(b) Lava unit with narrow spinifex-textured zone; zones as in (a) but at the top of the B zone there may occur schlieren of glass with microspinifex texture.

(c) Differentiated komatiite flow.

1. Flow top breccia lies on top of glassy zone with downward coarsening spinifex texture; at the base the spinifex matrix becomes crystalline and gabbroic, grading into 2.
2. Main gabbro unit, often glomerophytic and downward coarsening, grading down into 3.
3. Feldspathic pyroxenite unit which may have cumulate textures.
4. Oikocrystic peridotite consisting of massive peridotite with pyroxene oikocrysts enclosing olivine cumulate; this grades downward into 5.
5. Pyroxene-rich peridotite which does not contain oikocrysts.
6. Basal melanogabbro with a basal glassy layer with microspinifex.

(a) and (b) are based on flow units at Munro Township, Ontario, Canada; after Arndt, quoted in Donaldson (in Arndt and Nisbet, 1982); section (a) may exceed 10 m; (b) would be typically 10 m; (c) is from the Cape Smith fold belt, Quebec, Canada; after Hynes and Francis (in Arndt and Nisbet, 1982); the section is thicker than 100 m.

relationship. The REE and LIL (large-ion-lithophile element) data suggest a source similar to, but more depleted than that of tholeiitic basalts. Thus although no consensus currently exists, most proposed models concentrate on a depleted (MgO-enriched) upper mantle or subducted ocean plate source region. Either fractional melting to a larger degree than necessary to generate basalt, or further melting of a harzburgite residue left after basalt generation, are favored. Other proposals consider the possibility of "wet" lherzolite in the Archaean mantle, or emphasize the role of high Archaean geothermal gradients, which apparently explain the declining abundance of komatiites with time. A review of various models and extensive reference lists may be found in Allegre (1982—see Arndt and Nisbet 1982) and Tarney and Windley (1981).

Economic aspects

Since the late 1960s komatiites have assumed an increasing importance as a potential source of Cu and Ni. The association of sulfide mineralization with these rocks was recognized long before their true nature was understood, especially in Canada, where the relationship between "bird's feet" or "chicken track" peridotites and Cu-Ni-bearing Fe-sulfide deposits was recognized in the 1920s and 1930s.

Disseminated sulfide, usually a mixture of chalcopyrite, pyrite, pyrrhotite, and pentlandite is widespread in komatiites, but is rarely economic. However, in a few localities massive segregations of sulfide occur at the base of massive flows (e.g., Kambalda, W Australia; Munro Township, Ontario, Canada), usually near the bottom of the lava pile. The source of the sulfur has proved highly controversial, although the mechanism by which sulfide droplets concentrate Ni particularly in an ultramafic melt has proved less so (Usselman et al., 1979; see review by Naldrett and Campbell in Arndt and Nisbet, 1982).

ADRIAN F. PARK

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Cross-references: *Archaean volcanism: Flood basalt; Greenstone; Igneous rocks—tectonic setting; Ophiolite; Spinifex texture; Ultramafic rocks.*

ULTRAMAFIC ROCKS

Ultramafic rocks (*ultramafites* or *ultramafolites*) are igneous and metamorphic rocks that consist predominantly of mafic (ferromagnesian) and related minerals. The more important primary mafic minerals are olivine, pyroxene, hornblende, mica, garnet, and opaque ores; the most abundant secondary or metamorphic minerals are serpentine, chlorite, amphibole, epidote, and magnetite. The principal primary minerals occur in almost all proportions, and monomineralic rocks are common.

The International Union of Geological Sciences Subcommittee on the Systematics of Igneous Rocks (Streckeisen, 1976) restricted the term "ultramafic" to rocks that contain <10% plagioclase, but others (Wyllie, 1967) include as much as 30% felsic minerals. These differences arise from the fact that ultramafic rocks grade into gabbroic rocks by increase in plagioclase or feldspathoids and represent end-members of several igneous assemblages (Table 2).

The names and mineralogical compositions of the more common varieties of ultramafic rocks are shown in Table 1 while the following are less common types (or terms).

Afrikanidite—ultramafic rock composed of knopite, titanomagnetite and melilite (Afrikanda, Kola Peninsula, U.S.S.R.).

Amphibolide (*amphibololite*)—a field term used to designate any coarse-grained, holocrystalline igneous rock composed almost entirely of amphibole minerals.

Anorthitissite—a hornblendite containing anorthite.

TABLE 1. Mineral composition of principal kinds of ultramafic rocks

		Olivine	Orthopyroxene	Clinopyroxene	Chromite, spinel	Hornblende	Garnet	Plagioclase	Biotite, phlogopite	Ilmenite	Magnetite	Feldspathoid
Peridotite	Dunite	P	a	a	a			a		a	a	
	Harzburgite	M	M	a	a	a				a	a	
	Lherzolite	M	M	M	a	a	(s)	a		a	a	
	Wehrlite	M	a	M	a	a	(s)	a		a	a	
	Chromitite	M	a	a	M			(M)				
Pyroxenite	Orthopyroxenite	a	P	a	a					a	a	
	Websterite	a	M	M	a	a	(a)	a		a	a	
	Clinopyroxenite	a	a	P	a	a	(s)	a		a	a	
	Hornblendite	a	a	a		M		s	a	a	a	
	Kimberlite	M	M	M	a		s		s	s	a	
	Eclogite	a	a	M			M			a	a	
	Jacupirangite, melteigite etc.	a		M				a	s	s	M	a

P, predominant—>90%; M, major—20–90%; s, subordinate—0–20%; a, accessory—<10%; () indicates uncommon but petrologically important occurrence.

Avezacite—an augite–hornblende–titaniferous Fe-ore igneous rock with a cataclastic structure (Avezac, Pyrenees).

Biotitite—an igneous rock composed almost entirely of biotite.

Cumberlandite—abundant olivine in a ground-mass of ilmenite, magnetite, labradorite, and accessory spinel (Cumberland, Rhode Is., U.S.A.).

Davainite—a hornblendite,? metamorphic.

Gronlandite—a hypersthene hornblendite.

Kvellite—a dark-colored rock with lamprophyric affinities and phenocrysts of lepidomelane, olivine, barkevikite, apatite, ilmenite, and magnetite.

Lherzite—a hornblendite composed mainly of brown hornblende with minor proportions of biotite and ilmenite and occasional garnet; the heteromorphic form of theralite.

Pallasite—any ultramafic rock, whether of meteoritic or terrestrial origin, that contains ca. 60% Fe in the former or more Fe-oxides than SiO₂ in the latter.

Pedrosite—a hornblendite (Pedrosa, Portugal).

Perknide—any holocrystalline rock composed almost entirely of dark-colored minerals (e.g., amphibolite, peridotite, pyroxenite).

Perknite—rocks composed essentially of pyroxenes or amphiboles; ultramafic rocks is a more commonly used term.

Sörkedalite—a feldspathoid-free ultramafic rock resembling essexite or kjelsasite in

composition, but with high Ti, Fe, and P contents.

Classification

Classification on the basis of the kinds of associated rocks and geologic occurrence has been used as a means of comparing and contrasting ultramafic types, with Wyllie (1967, 1970) recognizing 11 types, including metamorphic derivatives. Salient features of seven types are outlined in Table 2. The principal features of stratiform, concentric, and alpine-type ultramafic masses and their relationships to associated rocks have been summarized by Jackson and Thayer (1972). The alpine-type peridotites are best-known for their association with gabbro and basaltic volcanic rocks in the ophiolite assemblage, which is considered to represent oceanic crust and upper mantle. The komatiite assemblage (Arndt and Nisbet, 1982) comprises peridotite bodies, peridotite–gabbro complexes, and basaltic lavas, and is found in greenstone belts.

Kimberlites and alkalic ultramafic rocks are closely associated in stable platform areas. Kimberlitic rocks occur as solitary units and in carbonatites, and both assemblages comprise rocks that experimental data indicate were formed at high pressure deep in the mantle (Dawson, 1980). Eclogites are found as xenoliths in kimberlite, with alkalic rocks in

TABLE 2. Features, occurrence, and petrologic associations of some ultramafic igneous rocks

	Internal features and form of body	Associated rocks	Structural relations and tectonic setting
Stratiform association	Cumulus or settled textures, layered; tabular parallel to layers or funnel shaped	Gabbroic and anorthositic to granitic or alkalic	Intrusive into metamorphic or volcanic terranes
Concentric association	Cumulus, flowage, and tectonite fabrics; cylindrical, rudely concentric, irregular	Gabbro, tonalite, diorite, augite syenite, granodiorite	Intrusive in orogenic belts; late syntectonic to post-tectonic
Alpine association	Tectonitic, recrystallized, some relict cumulate texture; modified cumulate to metamorphic-type layering; irregular form and internal structure	Gabbro, troctolite, Na-rich granophyric rocks; basaltic to keratophyric volcanic rocks of the ophiolite assemblage	Tectonic blocks and diapiric masses in orogenic belts; from the upper mantle
Komatiite association	Porphyritic, cumulate, fluid mush, spinifex-type platy olivine; sheets, flows, dykes	Stratiform peridotite–gabbro complexes, Mg-rich basalts and tholeiitic pillow basalts	Flows, subvolcanic sheets(?) and other intrusions in Precambrian greenstone belts
Kimberlitic association	Various: porphyritic massive to fragmental and tuff-breccia; funnel-shaped or irregular diatremes, dykes, and sills	Alone or associated with carbonatites and alkalic complexes	Intruded along deep-seated tension fractures in stable platform areas
Alkalic associations	Various: cumulate, porphyritic to equigranular; cumulate layering, flowage features; complex intrusive relations; xenolithic masses	Gabbroic and syenitic rocks rich in alkalic feldspar and feldspathoids; carbonatites	Intrusive ring complexes, small stocks, dykes, sheets; in stable platform areas, subvolcanic
Inclusions in volcanic rocks	Cumulate, igneous and tectonite fabrics; xenoliths in cinder cones, diatremes, and basalt flows	Wide variety of rocks, from dunite and eclogite to crustal rocks; most abundant in alkalic basalts	Arcuate volcanic belts; continental circum-oceanic, oceanic, and intracontinental

diatremes and, much less abundantly, with other ultramafic nodules in alkalic basalts. Nodules in basalts appear to be much more closely related to the alpine-type peridotites than to any other type; they are found in alkalic basalts in continental circum-oceanic or intra-oceanic volcanic belts.

Most ultramafic metamorphic rocks are formed by alteration of peridotite or pyroxenite to serpentinite or talc–amphibole–chlorite rocks that commonly are schistose or sheared. Small boudin-like ultramafic masses, however, between a few centimeters and about 1 km in size, which occur in amphibolite and pyroxene granulite, are considered to have been formed by metamorphic differentiation. During recrystallization at high temperatures and pressures, pyroxene- and olivine-rich banded rocks appear to have been formed, possibly from impure dolomite or serpentinite; in the same way hornblende-rich bands are formed in gneisses from massive basaltic rocks. Formation of large ultramafic bodies in this manner seems unlikely (Sørensen, 1967), but some Russian petrologists have proposed a metasomatic derivation for olivine-rich peridotite and dunite (Moskaleva, 1966).

Textures and structures

Ultramafic rocks show a much greater variety of textures and structural features than any comparable group of rocks because they have formed under conditions that range from volcanic to those deep in the mantle. They may be formed in two very different ways—by crystallization from liquid magma and as residues from partial fusion. Formation of basaltic magma and peridotite simultaneously by partial melting of mantle material and removal of the melt is a widely accepted corollary of the theory of ocean-floor spreading. Evidence of partial fusion to form liquid magma of gabbroic or pyroxenitic composition plus a solid residue of olivine-rich peridotite or dunite has been found in many inclusions in alkalic basalt and in alpine peridotites of the lherzolite subtype (Boudier and Nicolas, 1972).

Because settling of early-formed mafic minerals almost invariably accompanies crystallization of quiescent basaltic or more mafic magma, many ultramafic rocks are characterized by primary textures and structural features that resemble those of sedimentary rocks. Such cumulate textures and structures, including

graded layering, cross-bedding, and mushy-rock slumping, are highly developed in the lower parts of some large stratiform complexes (see **Bushveld Complex**; **Skaergaard intrusion**). Primary cumulus features are commonly most obvious in transitional facies between ultramafic and gabbroic rocks. The most complete crystal settling probably is shown by spinifex texture in peridotitic lava flows. In these flows, plates of olivine apparently grew downward from the surface in peridotitic liquid from which even minute skeletal olivine crystals had settled when the lava was ponded (Pyke et al., 1973). Differentiation by crystal settling in the upper mantle, just as in intrusions in crustal rocks, is shown by relict cumulus features in many alpine peridotites and ophiolitic complexes (Jackson and Thayer, 1972). Various noncumulus igneous textures occur in minor intrusions within layered complexes or alkalic ultramafic rocks.

Alpine-type peridotites and eclogites have metamorphic fabrics and secondary structures that have been imposed on primary features by solid or quasi-solid flowage at high temperatures. Most alpine-type peridotites show evidence of high-temperature deformation, and in some lherzolites this deformation is accompanied by partial fusion and re-equilibration from high-pressure to lower-pressure mineralogy. Ultramafic nodules in basalt show a similar complete range from undeformed cumulate textures to re-equilibrated and completely recrystallized tectonites. Kimberlites are marked by unusual "porphyritic" textures that grade into breccias. In the more massive facies, megacrysts appear to have reacted with a fine-grained groundmass that is rich in hydrous minerals. In the breccias, megacrysts appear to have been abraded during solid movement by gaseous fluidization mechanisms.

Geochemistry

Although limited by the predominance of mafic minerals, chemical variation in ultramafic rocks ranges between the compositions of the monomineralic rocks listed in Table 1 and may be substantial. Alpine-type peridotites of the harzburgite subtype are the most uniform: olivine and enstatite range from ca. 88 to 94% Mg_2SiO_4 and MgSiO_3 , respectively; SiO_2 varies with the ratio of olivine to pyroxene; CaO is low because enstatite is the dominant pyroxene; and Al_2O_3 is less than 2.5%. The lherzolite subtype may contain 1.5–5% Al_2O_3 in aluminous pyroxene or garnet. Alpine-type peridotites are notably deficient in Na, K and Ti and low in Fe. In stratiform ultramafic rocks, olivine and orthopyroxene range from ca. 75 to 94% of the Mg end members, but CaO and Al_2O_3 are

higher than in most alpine-type peridotite because of interstitial plagioclase and clinopyroxene. In the concentric complexes, the rocks range from dunite, in which olivine is as much as 93% forsterite, through wehrlite, to hornblende pyroxenite, which is rich in Al_2O_3 and titaniferous magnetite, contains 2–2.5% $\text{Na}_2\text{O} + \text{K}_2\text{O}$, and is critically undersaturated with respect to SiO_2 .

T. P. THAYER

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Cross-references: *Alkaline rocks—undersaturated*; *Bushveld Complex*; *Cumulate textures*; *Diatreme*; *Eclogite*; *Greenschist facies*; *Greenstone*; *Harzburgite*; *Igneous rocks—classification and nomenclature*; *Kimberlite*; *Lamprophyre*; *Lherzolite*; *Mafic and ultramafic inclusions in igneous rocks*; *Ophiolite*; *Peridotite*; *Plutonic rocks—classification and nomenclature*; *Pyroxenite*; *Serpentinite*; *Skaergaard intrusion*; *Ultrabasic igneous rocks*; *Ultramafic belts*; *Ultramafic lavas*.

UNITS AND CONVERSION FACTORS

The units in the standard CGS system (centimeters, grams, seconds, calories) and in the SI system introduced in 1974, and the relevant conversion factors, are set out in McBirney, A. R., 1984, *Igneous Petrology*, San Francisco: Freeman, Cooper, pp. 464–465.

UPPER MANTLE—EXPERIMENTAL PETROLOGY

Knowledge of the Earth's upper mantle (*asthenosphere*) is derived from three major sources: (1) studies of natural rocks and magmas reasonably deduced to be derived from the upper mantle, (2) determination of physical properties of the upper mantle by geophysical techniques, usually yielding averaged or regional characteristics, and (3) laboratory reproduction of physical conditions existing in the upper mantle and determination of the mineralogy, petrology, and physical properties of potential upper mantle materials under relevant physical conditions. In the third area, routine usage of piston-cylinder high-pressure apparatus to subject reasonably large (10–30 mg) samples to closely reproducible pressures and temperatures up to 50 kb and 1,700°C, with, in addition, variation and control of other variables such as H_2O , CO_2 , and oxygen fugacity, has resulted in great advances in knowledge of the upper mantle. Other types of high-pressure apparatus have been used to attain pressures of up to 500 kb at temperatures $\geq 1,000^\circ C$, for syntheses of high-pressure polymorphs of common rock-forming silicates. This research has its principal application in understanding the transition zone of the upper mantle, at depths of ca. 300–1,050 km.

Experimental petrology studies may utilize single-, two- or three-component synthetic systems, complex multicomponent synthetic mixtures (see **Pyrolite**), or natural rocks and minerals. All have advantages for specific purposes. In determining the results of experiments, optical and X-ray diffraction methods are normally used and, for studies of complex, multicomponent systems and natural rock compositions, the use of the electron probe microanalyzer has, in recent years, become mandatory.

Subsolidus mineralogy

A commonly accepted model for the upper mantle is lherzolite composed of olivine (ca. For_{90}), Ca-poor orthopyroxene, Ca-rich clinopyroxene, and an aluminous phase such as plagioclase, spinel, or garnet. Models of this type are consistent with seismological data and are principally derived from studies of natural peridotites of high-pressure mineralogy included in kimberlites or alkaline basaltic magmas and as lenses in high-pressure metamorphic terranes. Thus, experimental studies have emphasized the subsolidus phase equilibria for lherzolite compositions, studying reactions between the major phases and varying mineral

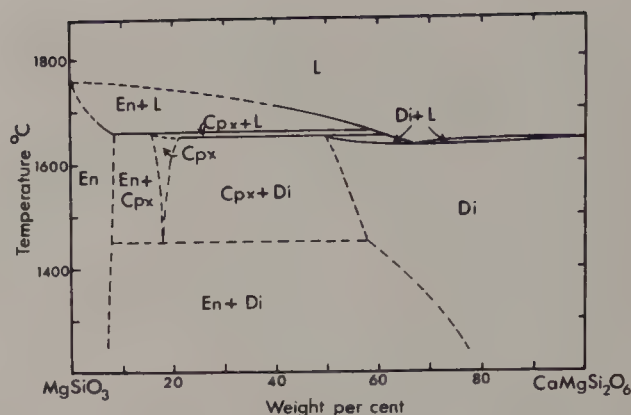


FIGURE 1. Experimental determination of phase relations at 20 kb load pressure between enstatite (En), subcalcic clinopyroxene (Cpx), Ca-rich clinopyroxene (Di), and liquid (L).

compositions, particularly of the pyroxene solid solutions, as functions of pressure and temperature.

The phase relations of the two pyroxenes synthesized in the system $MgSiO_3$ – $CaMgSi_2O_6$ as a function of temperature at 20 kb, and of pyrope and aluminous enstatite in the system $MgSiO_3$ – Al_2O_3 as a function of pressure are illustrated in Figs. 1 and 2, respectively. In

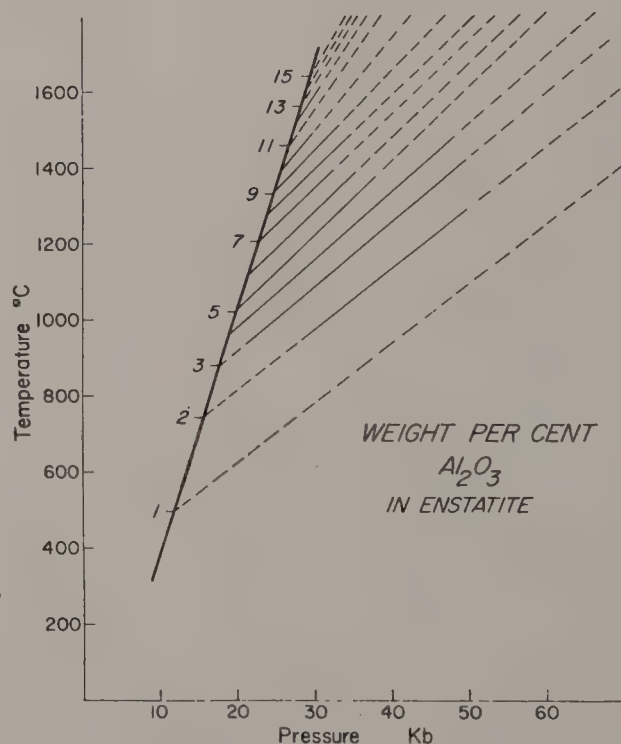


FIGURE 2. Experimental determination of the solubility of Al_2O_3 in enstatite, coexisting with pyrope garnet in the system $MgSiO_3$ – Al_2O_3 , as a function of pressure and temperature; garnet is stable with aluminous pyroxene to the right of the heavy line; numbers on that line refer to contours for wt. % Al_2O_3 in enstatite.

order to apply these data to natural rocks, i.e., to use the solid-solution gap between coexisting pyroxenes and the alumina content of orthopyroxene coexisting with garnet to estimate T, P conditions of crystallization of the rock, some extrapolations and simplifying assumptions are necessary. However, the diagrams have a basic importance in illustrating the type of relationship to be expected in the complex system and also in calculating basic thermodynamic data for the phases involved. These approaches lead to the ability to determine conditions of crystallization of natural mantle samples (Boyd, 1973).

The stability fields of plagioclase lherzolite, spinel lherzolite, and garnet lherzolite for several anhydrous, complex compositions suggested as models for the upper mantle are shown in Fig. 3. All natural assemblages contain olivine, orthopyroxene, and clinopyroxene and an aluminous phase that varies from plagioclase, through spinel (with Al-rich pyroxenes, particularly at higher temperatures), to garnet with increasing pressure. The differences in the positions of phase boundaries are attributable, at least in part, to the different bulk compositions studied, but are also due to experimental uncertainty.

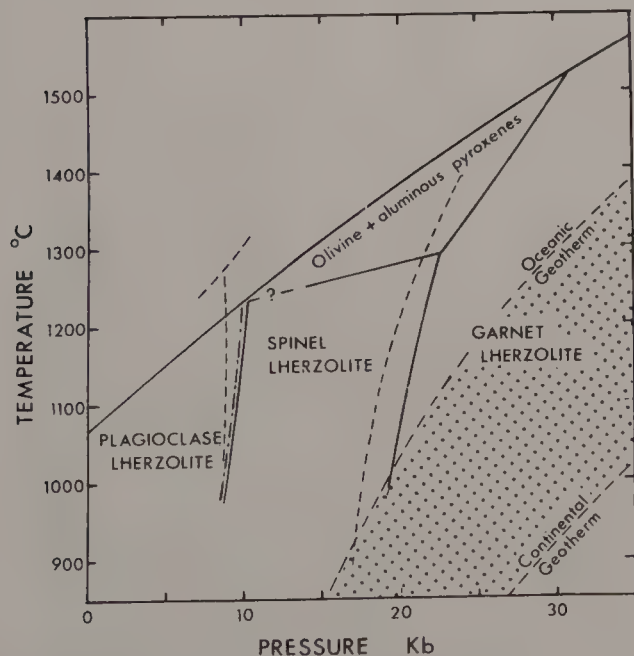
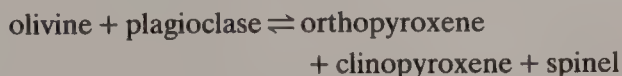
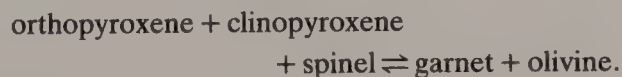


FIGURE 3. Stability fields of subsolidus mineral assemblages in lherzolite compositions; solid lines refer to boundaries determined for pyrolite composition, the dashed boundary for garnet appearance refers to data for mixtures of natural lherzolite minerals by O'Hara et al. (1971), the dot-dash boundary is from Emslie (1970) for reaction between intermediate plagioclase and Fe-bearing olivine, and the dashed boundary at 8 kb is from Herzberg (1972) for the anorthite + olivine reaction.

The boundary between the plagioclase lherzolite and spinel lherzolite facies is determined by the reaction



and the boundary between spinel lherzolite and garnet lherzolite facies by the reaction



If the Al_2O_3 /(normative pyroxene) ratio of the bulk composition is low (e.g., Al_2O_3 /pyroxene (wt.%) = 0.16 for "Hawaiian" pyrolite), all Al_2O_3 may enter the pyroxene at high temperatures between 10 kb and 30 kb yielding mineral assemblages of olivine + aluminous pyroxenes (Figs. 3, 5).

Figure 3 may be used to predict the nature of mineralogical zoning in the upper mantle by noting the depths of intersection of proposed geothermal gradients with the phase fields established by experimental petrology. If the upper mantle is anhydrous, Fig. 3 shows that the estimated geothermal gradients do not intersect the solidus and do not predict a partially molten layer beneath either stable continental crust or ocean basins.

Within the transition zone of the upper mantle (i.e., ca. 400–1,050 km) seismic velocities increase rapidly with depth, but most of this velocity increase is concentrated in two or three narrow zones around 400 km, 700 km, and possibly 1,050 km. These regions of rapid increase in seismic velocity are commonly referred to as seismic discontinuities and it has been inferred that the seismic discontinuities are caused by phase changes or polymorphic transitions in which the major silicates of the upper mantle (olivine, pyroxenes, garnet) transform to higher-density polymorphs characterized by more close-packed crystal structures. Techniques in experimental petrology that enable the attainment of pressures to 500 kb at temperatures exceeding 1,000°C have yielded syntheses of many high pressure polymorphs of rock-forming silicates or of their isostructural germane counterparts (Fig. 4). These experimental studies indicate that the rapid increase of seismic velocity around 350–450 km results from transformation of olivine to spinel-like structure (β phase) and from transformation of pyroxene structure to garnet structure with some silicon in octahedral coordination (Fig. 4). Understanding of the probable cause of the deeper seismic discontinuity at around 700 km relies largely on the studies of germane counterparts to the silicates, on solid-solubility relations of silicates in

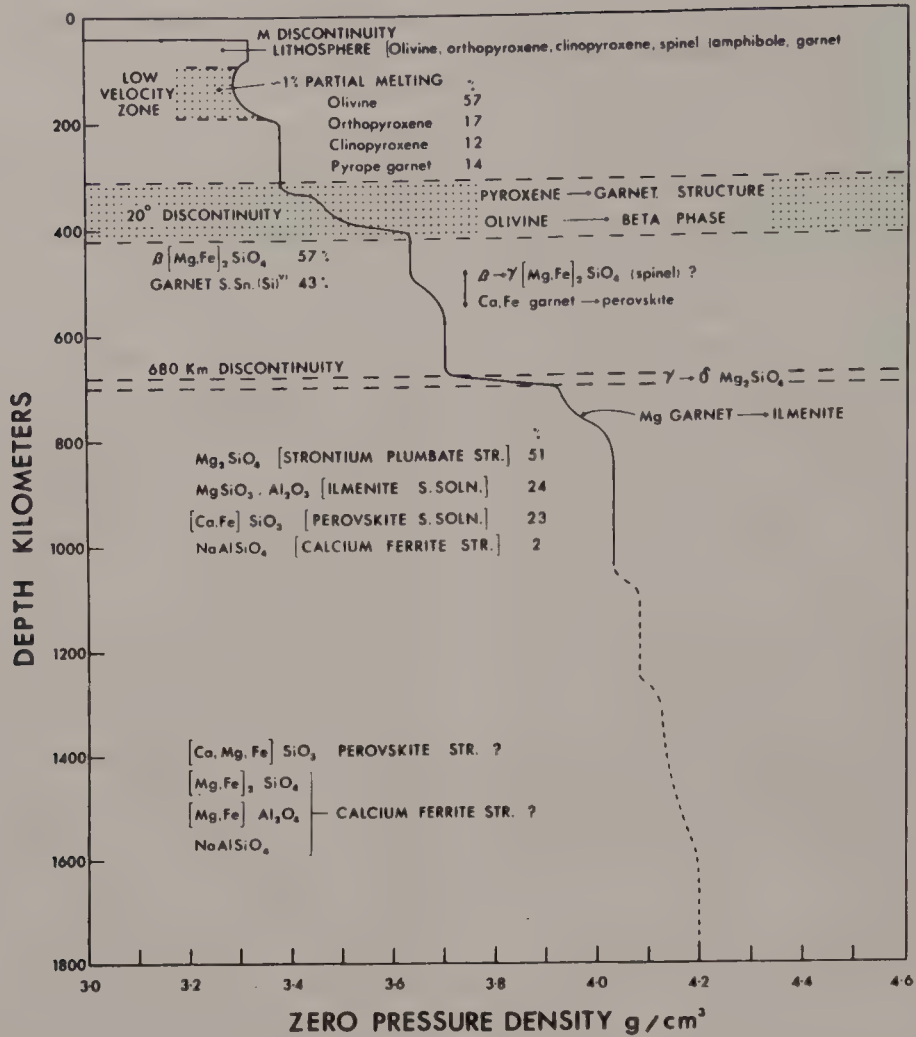


FIGURE 4. Depth intervals in the upper mantle at which polymorphic transformations of the major silicates occur and the density increase associated with the polymorphic transformation. (From Ringwood, 1975.)

high-pressure germane polymorphs on shock-wave investigations. It is probable that transformation of garnet structures to dense perovskite or ilmenite-type structures, and of $(\text{Mg, Fe})_2\text{SiO}_4$ spinel structure to a polymorph of strontium plumbate type or to simple close-packed oxides (periclase + stishovite) are responsible for seismic discontinuities in the 600–700 km interval.

Magma genesis in the upper mantle

There is compelling geophysical, petrologic, and geologic evidence that basaltic magmas are derived from the Earth's mantle by a partial melting process. Thus, since about 1960, there have been an increasing number of experimental investigations of the melting behavior of both basalts and peridotites under high-pressure conditions. These experimental results have led to several petrogenetic schemes for magma genesis, one of which is presented in the following paragraphs, emphasizing the experimental basis for the petrogenetic model.

Figure 5 illustrates the experimentally determined solidus of "Hawaiian" pyrolite and summarizes the compositions of liquids and nature of residual phases at temperatures above the solidus (Jaques and Green, 1980). At 15–20 kb, small degrees of melting (<10%) produce nepheline-normative basaltic liquids in equilibrium with residual olivine, orthopyroxene, clinopyroxene, and aluminous spinel. With increasing temperature and degree of melting, clinopyroxene is eliminated from the residue and liquids become hypersthene-normative olivine tholeiite or tholeiitic picrite. If a constant degree of partial melting (say 30% melting) is considered then liquids in equilibrium with residual olivine, enstatite, and chromite, change from quartz tholeiite at <5 kb to tholeiitic picrite at ~20 kb. The diagram is contoured for normative olivine content of liquids as a function of pressure and temperature.

The presence of water in volcanic gases and in hydrous minerals in mantle-derived inclusions

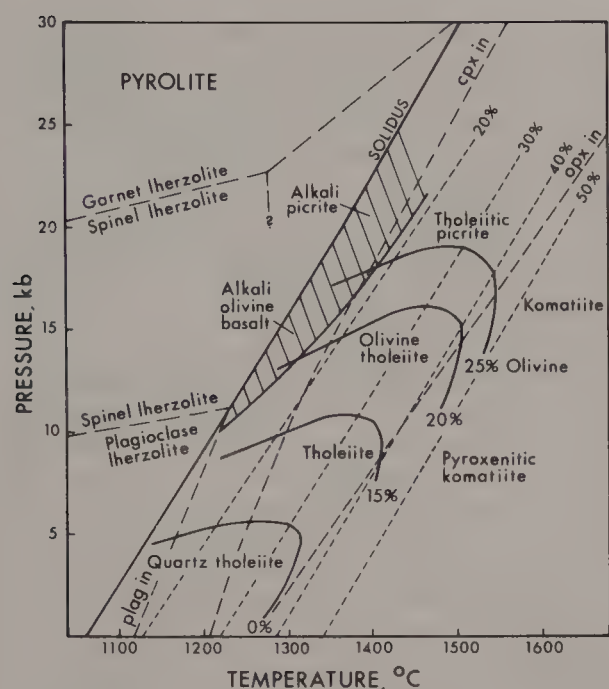


FIGURE 5. P - T diagram illustrating the products of anhydrous partial melting of "Hawaiian" pyrolite, showing contours of percentage liquid present (dashed lines), of normative olivine content (solid lines), and the presence or absence of residual phases (olivine and chrome spinel are residual phases for all conditions below 50% melting contour). (After Jaques and Green, 1980.)

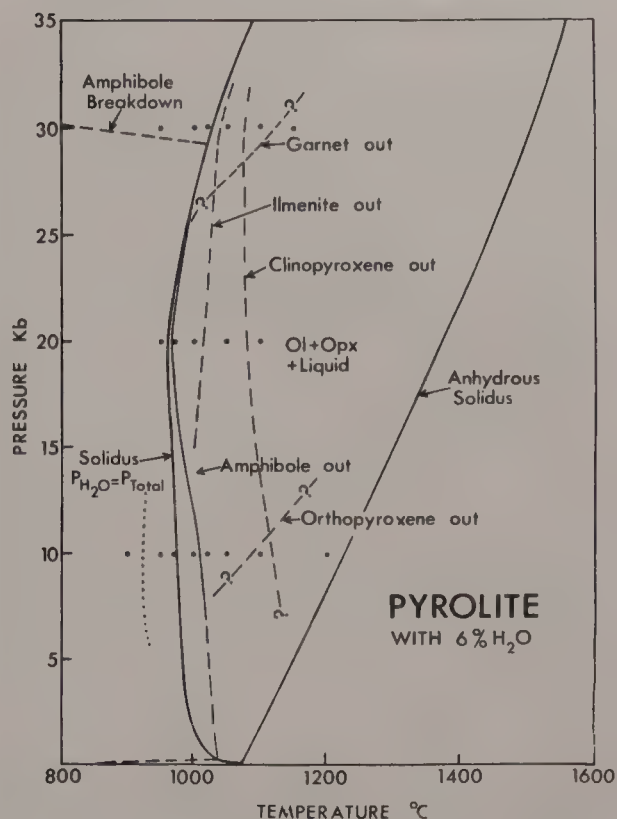


FIGURE 6. Experimentally determined solidus for pyrolite composition under water-saturated conditions and curves (dashed lines) showing the disappearance of phases above the solidus. (After Green, 1973.)

argues that water is a constituent of the upper mantle, and the presence of water greatly affects the position of the solidus and the nature of partial melt products (Kushiro, 1972). For water-saturated melting of pyrolite (Fig. 6), in contrast to anhydrous or very low P_{H_2O} conditions, orthopyroxene is eliminated before clinopyroxene at pressures up to 10–12 kb and liquids produced by partial melting of pyrolite under water-saturated conditions have higher SiO_2 contents than liquids produced by similar degrees of melting in the absence of H_2O or with $<0.5\%$ H_2O in pyrolite (see Fig. 8). At low to moderate pressure (up to ca. 20 kb) silica-oversaturated magmas may be produced by water-saturated melting, whereas if $P_{H_2O} \ll P_{TOTAL}$, the magmas at similar pressures but higher temperatures are olivine-normative. The main application of the results of experimental studies of water-saturated melting under upper mantle conditions, is to the understanding of magma genesis in the island-arc environment. In this region, the release of water from dehydration reactions (particularly amphibolite $\rightarrow$ eclogite), occurring within subducted crust and lithosphere in the Benioff zone, may cause high degrees of melting of peridotite above this zone,

producing hydrous magmas of island arc tholeiitic, calc-alkaline, or boninitic (high-Mg andesite) types. Differing major- and trace-element characteristics appear to reflect the variably refractory harzburgite to lherzolite of the lithosphere above the Benioff zone, the extent of large-ion lithophile element enrichment from the subducted slab, and the water content and condition of melting.

In contrast to this situation, the basaltic magmas of the mid-oceanic ridge or island-chain environments have very low water contents that imply water contents in their source region of $<0.5\%$ H_2O . If the water content of pyrolite is <0.4 – 0.5% , then at pressures up to 28–29 kb, 1,000°C, all water is held in pargasitic amphibole and the solidus for pyrolite has the form illustrated in Fig. 7. The sharp depression of the solidus at 27–30 kb is due to the instability of amphibole at high pressure relative to the denser assemblage olivine + pyroxenes + garnet + phlogopite + water (fluid). The phase diagram of Fig. 7 suggests a specific petrologic model for the lithosphere and low-velocity zone of the upper mantle. Estimated oceanic geothermal gradients in regions away from active spreading centers would intersect the

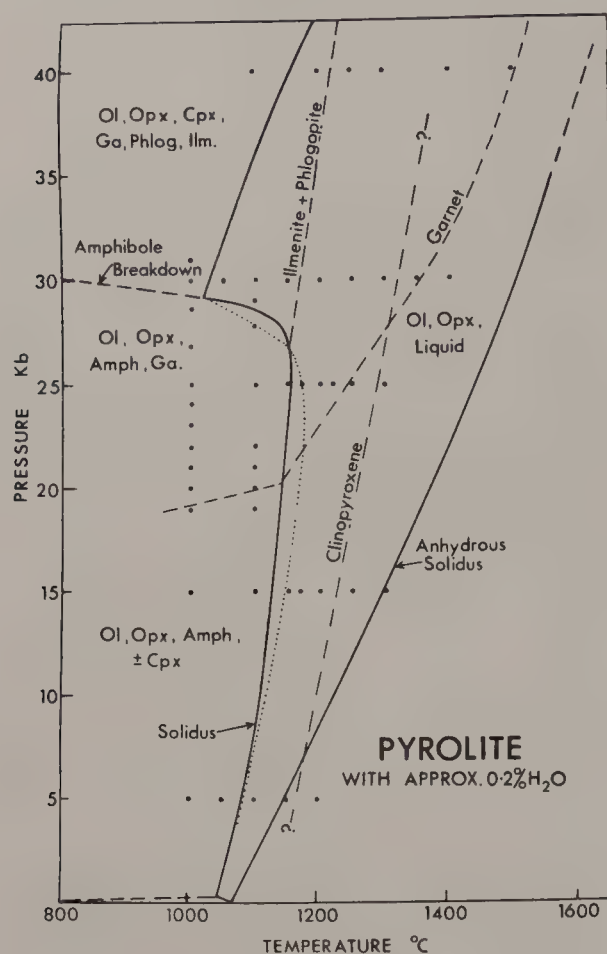


FIGURE 7. Experimentally determined solidus for pyrolite composition containing approximately 0.2% H_2O (i.e., $P_{H_2O} < P_{TOTAL}$) and all water in subsolidus at <30 kb held in pargasitic amphibole; dashed curves illustrate the sequence of disappearance of phases above the solidus.

pyrolite + ca. 0.2% water solidus at depths of 85–95 km (Fig. 7) and at deeper levels would pass into a region with a small (1–2%) melt fraction. The latter region is inferred to be the low-velocity zone, with the low seismic velocity and high seismic attenuation being attributed directly to the presence of the interstitial melt fraction. In regions of low geothermal gradient, such as beneath stable shield regions, the depth to the top of the low-velocity zone (i.e., the intersection of geotherm and solidus) will be greater and the degree of partial melting (and thus the decrease in seismic velocity and increase in seismic attenuation) in the low-velocity zone will be smaller. In regions of very high heat flow, the lithosphere will be thin and the high degree of melting present could lead to dynamic instability with diapiric upwelling of peridotite, increasing degrees of partial melting at shallow depths, and magma segregation—these are the conditions considered to be operative beneath active oceanic ridges or volcanic centers such as Hawaii or Iceland. The

detailed discussion of the applications of Fig. 7 illustrates the geophysical and tectonic implications that stem primarily from experimental determination of the phase relationships in a complex peridotite composition as a function of pressure, temperature, and water content.

In parallel with the experimental studies of lherzolite compositions, experimental studies of basaltic magmas, particularly of those that contain xenoliths of upper mantle origin and are reasonably deduced to be of upper mantle derivation, provide essential information towards building internally consistent petrogenetic models for magma genesis in the mantle. In general, the liquidus phase of olivine-rich basalts changes with increasing pressure from olivine, to orthopyroxene, to clinopyroxene, to clinopyroxene + garnet. As a specific example, the liquidus or near-liquidus phases of an olivine-rich basanite are olivine, orthopyroxene, clinopyroxene, and garnet at pressures near 30 kb and temperature of 1,250°C if the water content of the magma is 4% (approx.). The chemical compositions of these minerals are closely similar to those of residual olivine, orthopyroxene, clinopyroxene, and garnet present in pyrolite + 0.2% H_2O at 30 kb, 1,250°C. An internally consistent model predicts that the particular olivine-rich basanite could form at 30 kb, 1,250°C by 6% partial melting of (pyrolite + 0.2% H_2O) leaving a garnet lherzolite residue.

A similar study of extremely undersaturated (larnite-normative) olivine melilitite magma showed that dissolved carbon (as carbonate, CO_3^{2-}) was necessary to bring orthopyroxene onto the liquidus at pressures of 27–30 kb and for particular water (7–8%) and carbon dioxide (6–7%) contents, the olivine melilitite magma was saturated in olivine, orthopyroxene, clinopyroxene, and garnet at 27 kb, 1,160°C (Brey and Green, 1977). Experimental studies of this type, coupled with consideration of the major- and trace-element compositions of natural magmas, have led to a petrogenetic grid (Fig. 8) in which specific magma types and complementary crystalline residues are assigned depth and per cent melting parameters for a source of pyrolite + 0.2% H_2O composition.

Gabbro–eclogite transformation

In some metamorphic terranes, rocks of basaltic composition occur as the very dense (3.5–3.6) eclogite mineralogy (almandine–pyrope garnet + omphacitic clinopyroxene + quartz, kyanite, olivine or rutile). Laboratory studies of the mineral reactions involved in the transformation of gabbro to eclogite demonstrate the transitional role of mineral as-

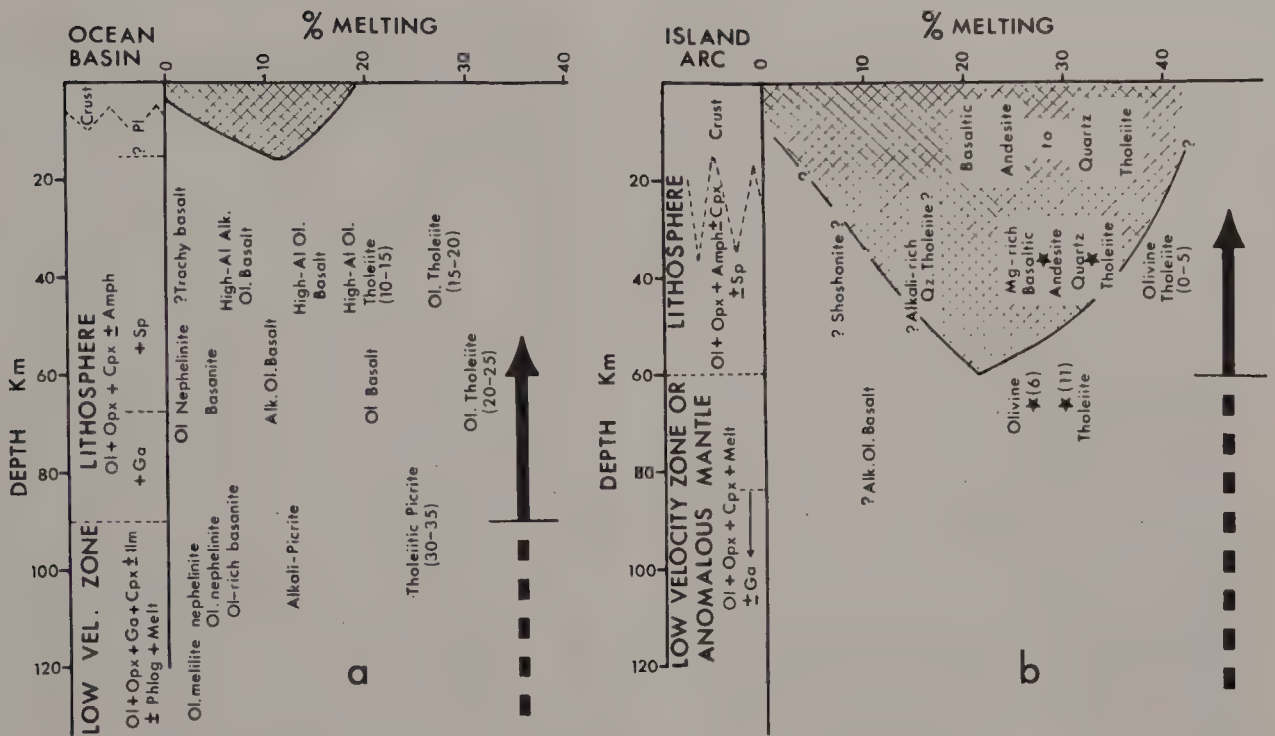


FIGURE 8. Summary diagram of petrogenetic applications of the melting studies on pyrolite composition showing the mineralogical character of the lithosphere and depth to the onset of partial melting in normal oceanic crust-mantle and island-arc situations. The character of magma derived by partial melting of pyrolite is plotted as a function of depth of magma segregation and degree of partial melting—numbers in parentheses adjacent to basalt names refer to the normative olivine content of the partial melt. Hatched area illustrates the range of conditions over which quartz-normative magmas may be derived by direct partial melting and magma segregation from pyrolite. (a) Compiled for melting under water-undersaturated conditions, with a water content in the source pyrolite of ca. 0.2% and with a small oxidized carbon content; (b) compiled for melting under water-saturated conditions. (From Green, 1973.)

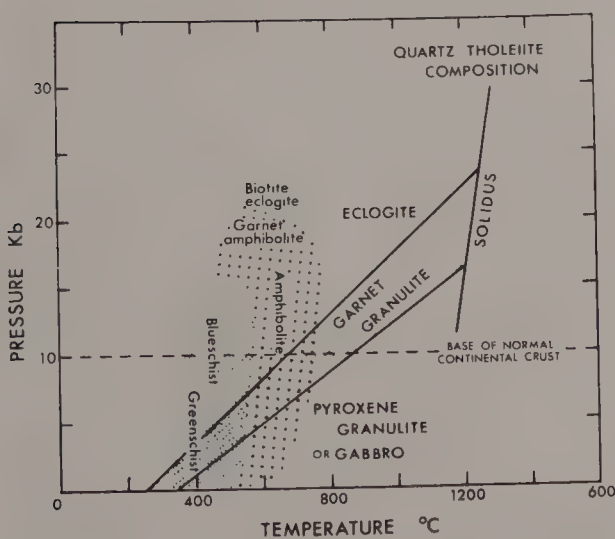


FIGURE 9. Experimentally determined stability fields of pyroxene granulite, garnet granulite, and eclogite for a dry quartz tholeiite composition, and probable relationship to the stability fields of hydrated mineral assemblages. (After Green and Ringwood, 1967a.)

semblages matching high-grade pyroxene granulites and garnet granulites and of the hydrated amphibolite (Fig. 9) assemblages. The experimental studies predict that eclogites are stable over a wide temperature range and that eclogites can form at quite low pressures (<10 kb) provided temperatures are below 600°C, the P_{H_2O} is very low, and kinetic factors are adequate to cause reaction at these temperatures. Experimental calibration of temperature-sensitive metamorphic reactions such as the stability of phengitic micas and the temperature-sensitive Fe, Mg partition between coexisting garnet and clinopyroxene, shows that natural eclogites do form over a temperature range at least between 600°C and 1,000°C.

From geologic and geophysical data, the nature of the oceanic crust, below thin sediment cover, is inferred to be of basaltic composition. Subduction of oceanic crust along oceanic trenches leads to, or may primarily be caused by, alteration of basaltic crust ($\rho = 3.0-3.2$) to eclogite ($\rho = 3.5-3.6$) with resulting gravita-

tional instability and sinking through the peridotite ($\rho = 3.3$) low-velocity zone. The P, T conditions under which the gabbro to eclogite reactions occur have thus been shown by experimental petrology to be appropriate to the P – T regime of a downgoing slab (subduction zone).

D. H. GREEN
T. MORI

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Cross-references: *Basalt*; *Eclogite*; *Gabbro and norite*; *Island arcs—petrology*; *Kimberlite*; *Lherzolite*; *Mafic and ultramafic inclusions in igneous rocks*; *Magma*; *Melilitite*; *Mid-ocean ridge and ocean-floor petrology*; *Peridotite*; *Phase transformation*; *Pyrolite*.

* Includes references in Fig. 3 explanation.

V

VOLATILES

Volatiles are those readily vaporizable materials present in a magma, such as H_2O and CO_2 , whose vapor pressures are sufficiently high that they will become concentrated in any gaseous phase that forms. Theoretically, all the components present in rocks can exist in a volatile phase under the appropriate conditions and in practice many of the so-called non-volatiles may exist as volatiles to a certain degree under conditions existing within the Earth. Although only a fraction of the total amount of these components may be volatilized, they should be considered as a portion of the total volatile phase and may be of great significance in several Earth processes. The importance of volatiles in magmatic processes is clearly recognized. In addition, sedimentary, metamorphic, and ore-forming processes are strongly influenced by volatiles and they are also important in hydrologic, atmospheric, and biological phenomena.

Constituents of high volatility

Magmatic volatiles. Certain constituents involved in Earth processes will be present predominantly in a gaseous phase, or one of relatively high fluidity, when such a phase can coexist with the liquid and solid phases. This group of constituents can be characterized as having high volatility. This "gaseous" phase will contain one or more of the following: H_2O , H_2 , CO_2 , SO_3 , SO_2 , S_2 , H_2S , CH_4 , NH_3 , N_2 , Ar , O_2 , HCl , HF ; other species may also be important. As magmas are generated, these species will be released from rocks and will enter the gaseous phase when their various solubilities in the melt are exceeded. These volatiles will be carried with the magma as it moves upward and will be released from volcanoes or forced out of crystallizing intrusions. Once released, they move toward areas of lower pressure and temperature and may mix with other volatiles present in rocks or in the atmosphere. This change in environment and interaction with the rocks through which they travel can greatly alter the original volatile-phase composition.

The conditions under which a volatile phase

can develop in a magma can be described thermodynamically. This approach shows that specifying such conditions for any particular case is a complex problem that requires consideration of pressure, temperature, melt composition, molar volumes, activities, etc. Once developed, the volatile phase will have a profound effect on the magma. Water and halogen gas pressures drastically reduce rock melting temperatures, and their loss from the system can cause rapid crystallization. Volatiles can play a major role in magmatic differentiation, collection of ore-forming materials, the heat budget of igneous bodies, fracturing of surrounding rocks, and emplacement. Volatiles increase rates of crystal growth, diffusion, and reaction and, in general, help to attain equilibrium. The release of volatiles from magmas can result in powerful disruptive phenomena such as volcanic explosions with their attendant propulsion of solid and liquid matter and disruption of surface features. The decapitations of Krakatoa, Katmai, and Mt. St. Helens are outstanding examples. Release of volatiles at depth from crystallizing magmas may develop great pressures that can fracture rocks and form breccia pipes and diatremes. The diamond pipes of Africa may owe their emplacement to an increase of volatile pressures that propelled the pipe-filling material to the surface of the Earth from depths of well over 100 km.

The actual composition of the magmatic volatile phase is a subject of much debate. Several approaches can be used to gain information on compositions: (1) volcanic gas collections, (2) fluid inclusions, (3) mineral parageneses, and (4) mineral compositions. The evidence available points to the dominance of H_2O , CO_2 , SO_2 , and halogen gases; however, possible compositions fall within wide limits. Furthermore, less-abundant volatile species may have an influence far exceeding their small proportions through chemical effects on crystallization sequences, mineral compositions, and total magmatic history. Effects on physical properties of magmas such as viscosity, heat capacity, density, total volume, and ionization are also very significant. Owing to the lack of accurate knowledge about compositions, however, a quantitative estimation of the total effect

is not available. Experimental studies of systems that include volatiles have shed much light on processes and have explained many phenomena. These investigations usually consider only one or two volatiles at a time (especially H_2O) and can only approximate natural conditions.

Volatiles in metamorphism. Rock metamorphism occurs in the presence of volatiles and volatiles may often be the dominant factor in determining the products of metamorphic reactions. The importance of volatiles was not fully realized until experimental, isotopic, and theoretical studies revealed their strong influence. Modern studies in the field and the laboratory have repeatedly emphasized this influence.

As sediments are deposited, volatile components are trapped. Surface volatiles such as water, atmospheric gases, and swamp and marsh gases will be held in the pores of sediments. Potential volatile components such as S, C, Cl, F, and Br may be dissolved in the pore fluid or included as part of the organic compounds and carbonate, evaporite, and hydrous minerals. These volatile components will react during burial and diagenesis to form new minerals and different pore fluids. Much of the pore-filling constituents will be lost during compaction, but significant amounts remain in rocks at depths up to many thousands of meters.

Increased metamorphic intensity leads to the formation of a series of volatile-containing minerals, e.g., zeolites, micas, and amphiboles. The higher grades of metamorphism favor a reduction in the volatile content of metamorphic minerals and expulsion of these components as a volatile phase. Several volatile species may be involved simultaneously. For example, water-bearing limestones release CO_2 in large amounts as well as pore-filling water. Carbonaceous shales release large amounts of H_2O from minerals and pore fluids and carbon and sulfur gases from organic compounds. Pore fluids are usually complex brines containing other minor volatile species.

Throughout these processes, volatile pressures such as P_{CO_2} , P_{H_2O} , P_{O_2} will exert an important or controlling influence on the minerals produced and reaction temperatures will be dependent upon volatile pressures as well as rock pressure. Experimental and theoretical studies have shown that complete changes in mineralogy can occur, at constant temperature and rock pressure, simply by varying the amount of volatiles or their pressures. Metamorphic mineral facies have classically been considered to be dominated by temperature variations, with rock pressure playing a secondary role. This new information has required a reinterpretation of metamorphic

facies in many cases. Field studies support the laboratory data.

Experimental studies of rock melting have shown that water pressure drastically reduces the melting temperatures of sediments and igneous rocks. Similar studies with other volatiles show additional strong effects. These results are important in determining the upper boundaries of metamorphic processes. Migmatites may represent partial rock melting and many earth scientists hold the view that many granitic bodies owe their origin to anatexis (melting of sediments). Metasomatism does occur on a limited scale and is thought by some earth scientists to be a large-scale process leading to major metamorphic zonings and the origin of granitic bodies. This process relies on the movement of volatiles through rocks and their ability to transport less-volatile species such as Fe, Mg, K, and Ca. The process does not readily lend itself to experimental investigation, largely because of the large time factor involved. However, extensive field and theoretical studies have provided strong evidence in its support and again have emphasized the importance of the volatile phase.

Volatiles in ore-forming processes. Many ore deposits owe their transport and deposition to volatiles. This process may be an extension of volatile-phase evolution from magmas or it may be a movement of material derived from adjacent rocks by lateral secretion. In either case, volatiles play a dominant role. The exact nature of the various ore fluids is in doubt; they may be gaseous, liquid, supercritical, or a combination of these. Water is the dominant volatile species in most cases, with CO_2 , sulfur gases, and halogens also important. The transport mechanism by which the ores are carried has been attributed to salt solutions, sulfur complexes, or volatility of the ore-forming constituents. The ore-forming fluid is a subject of intensive study that has resulted in a large body of information on the chemical and physical behavior of volatiles at elevated pressures and temperatures.

Constituents of low volatility

Constituents with low vapor pressures require special conditions of temperature and pressure to be present in the volatile phase in significant amounts. These constituents form a group that is distinct from the constituents of high volatility. However, the importance of low-volatility elements and compounds in the volatile phase makes their study a primary concern of many earth scientists. Among these constituents are the ones that form ore deposits

and volcanic and thermal area sublimates. These constituents may be present as volatile elements, halide compounds, and sulfur compounds and perhaps as volatile hydroxides and carbonates.

The existence of species of low volatility in a volatile phase requires the existence of either a supercritical fluid or a gaseous phase coexisting with a liquid phase. In the latter case, the volatile species must be present in a total amount that exceeds its solubility in the liquid under the prevailing conditions. The amount in the volatile phase will be a complex function of its interaction with other species and of its distribution between phases. The species will tend to establish equilibrium within the entire system. Therefore, the amount of a given species in the volatile phase depends upon the temperature, total pressure, and the amounts of all other constituents present as well as upon the ideal or non-ideal behavior of all the constituents. The description of a volatile phase of only five or six chemical components is a complex problem, while the approximation of a natural system with 10–20 significant components is a major problem. A large body of data exists on the solubilities, thermodynamic properties, vapor pressures, etc., of certain species of importance. Some data exist on the interaction between simple combinations of species. Conditions of higher temperature and total pressure, as well as more complex systems, must be studied further.

Much of the available information on the less-volatile constituents has come from field studies. These studies have been particularly fruitful when sublimates around volcanic fumaroles are investigated. Ore-deposit parageneses and mineral compositions are also valuable in estimating volatilities. The types of volatiles required to form various minerals can be determined and a thermodynamic approach can be used to establish limits on individual volatile-component pressures. These limits cover a broad range, however, owing to the inaccuracy of data and the assumptions that must be made. The results of studies on constituents of low volatility indicate that they may be often be present in the volatile phase in significant amounts. In some cases the paragenesis of minerals and the temperatures of deposition match the relative volatilities of compounds. Not all material transport can be attributed to volatiles, however, and some constituents of great importance in deposits are of such low volatility that such transport must be unimportant. Under subcritical conditions, the interplay of liquid and gaseous phases must be considered and solubilities in the liquid must account for the bulk of the transport.

Related terms

Fugitive constituent—a substance, such as a volatile constituent, originally present in magma, but lost during crystallization.

Juvenile water—gases and water that are derived directly from magma and that come to the surface of the Earth for the first time.

Phreatic gas—gases formed as the result of contact of ascending magma with surface or atmospheric water.

Plutonic water—juvenile water from magma at a considerable depth.

Resurgent—refers to magmatic gases or water derived from ground water, atmospheric water, or assimilated country rock (cf., phreatic gas).

BERT E. NORDLIE

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Cross-references: *Anatexis*; *Fumarole*; *Granites and mineralization*; *Greisen*; *Hydrothermal processes*; *Magma*; *Metamorphism—controls*; *Metasomatism*; *Pegmatite*; *Pneumatolysis*; *Volcanoes and volcanism*. In Vol. IVA: *Gases—volcanic*.

VOLCANIC BRECCIA

Classification and nomenclature

A *volcanic breccia* is a rock composed predominantly of angular fragments resulting from brecciation or emplacement due to volcanic action: it may, or may not, have a matrix. Four criteria have commonly been used in their classification: (1) cause of brecciation, (2) environment of deposition or emplacement, (3) location (in relation to geologic forms), and (4) rock type. Cause of brecciation has been used in cases where intrusion-breccia formed as the result of intrusion of magma into wall-rock (Fig. 1). Where there has been movement of the



FIGURE 1. Intrusion-breccia consisting of blocks of banded granulite in a matrix of granite situated in the deeper levels of a caldera structure; Lake Baikal region, Siberia, U.S.S.R.

brecciated material and the cause of brecciation is more difficult to establish, terms such as intrusive breccia indicate environment of development. Location, as reflected by such terms as vent-breccia and contact-breccia, can be avoided in most instances by using the appropriate term to describe the process of formation, e.g., vent-breccia is normally a pyroclastic breccia. Rock type can rarely be used on its own as sedimentary breccias, pyroclastic breccias, and deep-seated explosion-breccias may have comparable compositions.

Confusion has often been caused by misuse of noun-adjectives and adjectives to designate the criteria set out above (Wright and Bowes, 1963). Differences in meaning between these are demonstrated by the terms "intrusion-breccia" and "intrusive breccia". The former refers to breccia caused by intrusion of some other material, while the latter refers to a breccia which, as a mass, has cross-cutting (intrusive) relations, i.e., a noun-adjective is used to refer to the *cause of brecciation* (intrusion-breccia, explosion-breccia, etc.) while an adjective is used to refer to *environment or mode of emplacement* (pyroclastic breccia, hydrovolcanic breccia, intrusive breccia, etc.). The compound noun should not be separated by the insertion of an adjective but may be qualified by an adjective indicating subsequent geologic history: an intrusive explosion-breccia refers to a breccia formed as the result of (volcanic) explosive activity that shows intrusive relations to the surrounding rocks.

Adherence to this system of nomenclature would result in the use of self-explanatory terms that mean what they say. In addition misleading

terms, such as "explosive breccia", would be avoided.

Types

There are three types of rocks composed predominantly of angular fragments that have been fractured by volcanic agencies before deposition or emplacement.

Pyroclastic breccias. Rocks that are ejected from a volcano as a mass of large or small fragments, mostly of igneous parentage but with variable proportions of wall-rock, were defined as being pyroclastic by Wentworth and Williams (1932). Features of many variants are set out by Parsons (1969). These include vulcanian breccias associated with aerial ejection during explosive eruptions, breccias associated with strombolian and lava-fountain eruptions, pyroclastic flow-breccias, hydrovolcanic breccias related to phreatic eruptions, laharic breccias, vent-breccias and many types of breccia formed under water or under ice (see **Explosive volcanic eruptions—classification; Volcanoes and volcanism**). The terminology of pyroclastic deposits, including pyroclastic breccias, is given by Wright et al. (1980).

Autoclastic and alloclastic breccias. Autoclastic breccias ("broken by themselves") include the products of fragmentation of semisolid lava by its own movement (*friction-breccia, flow-breccia*) and the products of explosion of gases within lavas. They are commonly well-expressed in viscous acidic lava such as rhyolite and many rhyolite flows are largely composed of autoclastic breccia.

Alloclastic (Greek, *allos* other, *klastos* broken) breccias were formed by the fragmentation of pre-existing rocks by volcanic-related processes beneath the surface, i.e., they are neither pyroclastic nor autoclastic (Wright and Bowes, 1963). *Intrusion-breccia*, a term introduced by Harker (1908) for the results of penetration of an older rock by a network of ramifying veins leading to the newer rock enclosing detached blocks of varying size (Fig. 1), occurs in plutonic as well as subsurface volcanic environments. The term *explosion-breccia* which was introduced by Tyrrell (1928), and redefined by Blyth (1940) with Tyrrell's approval, refers to a rock formed more or less in situ by localized subsurface volcanic explosions involving the shattering of country rock (Fig. 2a). This activity may be related to the structurally controlled uprise of volatiles ahead of volatile-rich rising magma (Bowes and Wright, 1967; Wright and Bowes, 1968) with variations in gas pressure related to explosive activity resulting in spectacular zoned and rhythmic variations in minerals in the crystal-

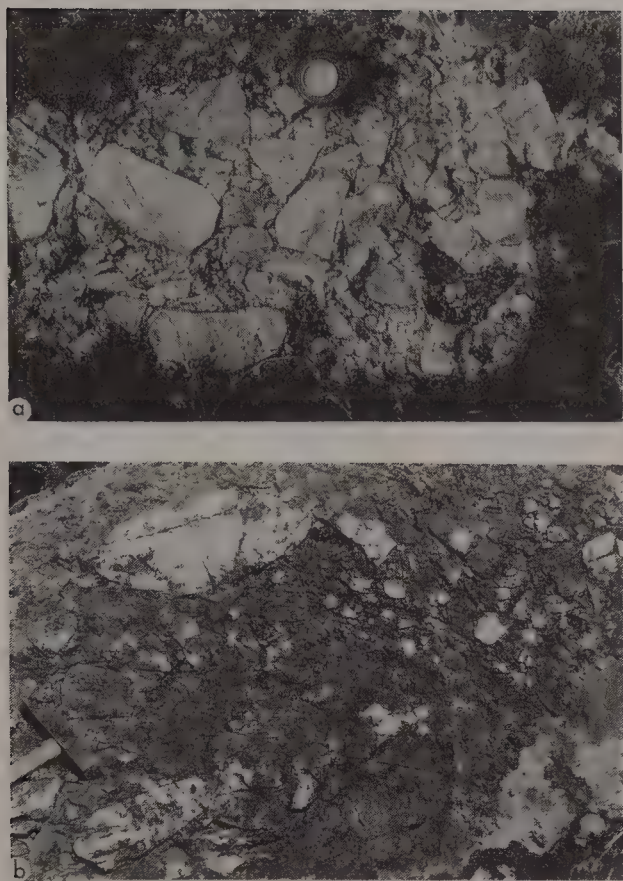


FIGURE 2. Explosion-breccia consisting of angular (a) and rounded with angular (b) blocks of quartzite in a matrix of comminuted quartzite situated in an explosion-breccia pipe, Back Settlement, Scotland (Bowes and Wright, 1961).

lizing magma (Hamidullah and Bowes, 1988; see **Appinite**; **Diatreme**). Movement of blocks of a breccia in rising streams of gases and/or magma can give rise to constituent blocks being rounded (Fig. 2b), faceted (Pitcher and Read, 1952) and having moved considerable distances, possibly associated with the process of fluidization (Reynolds, 1954). The resultant product is an *intrusive breccia*: the matrix may be tuffistic (*tuffistic breccia*), there may be no matrix following chemical and/or mechanical removal in escaping gas streams, or the matrix may be igneous. Some breccias similar in appearance to explosion-breccias, but with no associated igneous rocks, are heavily mineralized. The diatreme-facies of kimberlite occurs in explosion vents infilled with fluidized fragmental material.

There is considerable overlap between some of the types of volcanic breccia and some rock types may be derived from others—*intrusive breccias* may be explosion-breccias showing intrusive relationships. This is the result of the nature of the varied processes in volcanism in which brecciation, an integral part of all

volcanic activity, occurs at depth, near the surface, and at the surface.

D. R. BOWES

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Cross-references: *Appinite*; *Breccia pipe*; *Crypto-volcanic structure*; *Diatreme*; *Explosive volcanic eruptions—classification*; *Fluidization*; *Igneous intrusions—structural behavior*; *Kimberlite*; *Lava and lava eruptions*; *Pyroclastic eruption*; *Tuffisite*; *Volcanic neck*; *Volcanoes and volcanism*.

VOLCANIC GLASS

Volcanic glass is formed by rapid chilling of liquid lava when it comes into contact with the atmosphere, water, or cooler adjacent rocks. It occurs mainly as obsidian, pitchstone, and perlite in lava flows, domes, and dykes, and as pumice, cinder, and ash in pyroclastic rocks. The tendency of lava to form glass is a function of rate of cooling and chemical composition. Felsic lavas (those high in SiO₂, Na, and K)

tend to form glass more readily than mafic lavas (those low in SiO_2 and high in Mg and Fe). Thus, volcanic glass is more commonly found in rhyolites, rhyodacites, and dacites than in andesites and basalts but much, particularly, in old rocks has undergone *devitrification* with the development of a minutely crystalline texture. For general discussion of the properties of volcanic glasses see George (1924), Lacey (1959), and Murase and McBirney (1973).

Felsic glass

Obsidian, with its typical black vitreous luster and conspicuous conchoidal fracture, has long attracted the attention of man and has historically been used for artifacts—both weapons and art objects. Although it is most commonly black, greenish-black, dark gray, and reddish-brown varieties are known. Some obsidians are opaque; others show marked translucence. Black obsidian owes its color and opaqueness to the light-scattering effect of myriads of microscopic crystallites (embryonic crystals). The color of reddish-brown obsidian is due to oxidation of the small amount of Fe present. Greenish-black and gray varieties usually contain abundant minute vesicles (gas bubbles) that give them a silky luster. Obsidian of this type is particularly common in Mexico, where it is a popular material for artifacts.

Obsidian is rarely entirely free of crystals. Most contain a few tenths of a % of tiny crystals known as microlites (as distinct from crystallites, which are smaller and have no recognizable crystal form). Feldspar, pyroxene, and magnetite are the most common microlites. Volcanic glass that contains conspicuous macroscopic crystals is called *vitrophyre*.

Perlite is gray, pearly-lustered volcanic glass characteristically having an intricate system of concentric cracks resembling an onion structure. The cracking generally renders the glass quite friable. Perlite is usually associated with obsidian in nature and forms from it by the process of natural hydration (Ross and Smith, 1955). Obsidian contains only 0.1–0.3 wt. % H_2O , whereas perlite contains 2–4%. Isotopic analysis of hydrogen in the waters in these two types of glass indicates that the small amount of water in obsidian is of primary magmatic origin, whereas that in perlite is of meteoric origin (Friedman and Smith, 1958). Thus, obsidian alters to perlite by slowly absorbing meteoric water, and the perlitic cracks are produced by the internal strain that accompanies hydration. This discovery led to development of the obsidian-hydration dating method, which has had wide application in measuring the age of artifacts from archaeological sites (Friedman

and Smith, 1960). The chemical changes accompanying hydration include oxidation of Fe and decrease in Na (Lipman, 1965).

Pitchstone is black resinous-looking glass, usually associated with shallow intrusive rocks. It contains more water than does obsidian (commonly as much as 10%), the origin of which is not well understood.

Pumice is frothed volcanic glass formed usually during explosive eruptions by rapid expansion of contained gases, but it may form also by less-violent vesiculation at the surface of lava flows and domes. The porosity of pumice varies greatly but is commonly 50–75%. Some pumice is sufficiently porous to have a density of <1.0 and will float on water.

Volcanic ash is finely comminuted volcanic glass formed by extreme vesiculation of lava during explosive eruptions—ash is defined as ranging in size from 4 to 0.25 mm; smaller particles are generally called volcanic dust. Individual ash particles (*shards*) are usually curved, flattened, or cusped segments of vesicle walls, but some larger ash particles have fine cellular texture and are actually finely-vesicular pumice.

Pumice and ash may be carried by winds thousands of kilometers from their source volcanoes. Thin ash layers occur in sediments over much of the Earth's surface, in ocean-basin sediments, and they are also found in the Greenland and Antarctica ice caps. Such widespread thin ash and pumice deposits (tephra) are extremely useful in correlating stratigraphic units over long distances because commonly they are mineralogically distinctive and represent relatively short intervals of time.

Mafic glass

Mafic (basaltic and andesitic) lava tends to crystallize more readily than felsic lava, so that mafic glasses are much less common than felsic ones. Mafic glass also is more susceptible to alteration and weathering and so does not persist as long in time. Mafic glass occurs most commonly as scoria, cinder, pumice, and ash and in subaqueous pillow lavas and breccias.

Scoria is similar to pumice but is denser, with vesicles generally larger and more spherical in form, thus giving it a more openly cellular structure. Being rich in Fe it is usually red or brown owing to oxidation but may be black. Scoria and cinder are most commonly products of basaltic pyroclastic eruptions but also may form by vesiculation at the margins of lava flows.

Rare forms of mafic glass, produced by explosive vesiculation of very fluid lava, are "reticulite" or "thread-lace pumice" and "Pele's

hair." *Reticulite* is an interconnecting network of basaltic glass threads that has as much as 90% porosity. *Pele's hair* is basaltic glass that has been drawn into thin threads centimeters long but only a fraction of a millimeter thick. Both are produced in the hot updrafts of "fire fountain" type eruptions.

Pillow lavas and *pillow breccias* (Carlisle, 1963) form when basaltic lava pours into lakes, rivers, or the sea. Contact with water chills the outer surface of the lava, and oval or tongue-like "pillows" up to 2 m in diameter form. These pillows commonly have a thin rind of black basaltic glass (*tachylite* or *sideromelane*) and are commonly enclosed in a matrix of fragmented basaltic glass produced by fracturing and attrition of the pillow rinds during emplacement. Basaltic glass readily hydrates and oxidizes in wet environments to a yellow or brownish-orange mineraloid known as palagonite, and many pillow lavas and breccias consist predominantly of palagonite.

Other varieties

Amausite—a finely crystalline rock such as a devitrified glass.

Amerikanite—a natural glass once classed as tektite but now considered to be of volcanic origin (S America—Colombia, Peru).

Hyalomelane—commonly contains phenocrysts and differs from tachylite in its insolubility in acids; a basaltic vitrophyre.

Hydrotachylite—similar to tachylite but with a higher proportion of water (up to 13%).

Lassenite—originally considered to be trachytic in composition, but is dacitic; *metabolite* is the altered form (Lassen Peak, California, U.S.A.).

Marekanite—a variety of perlite obsidian containing masses of perlite that are clear glass and have a much lower water content than the surrounding perlite (Marekana, U.S.S.R.).

Okawaite—an aegirine-augite-bearing pitchstone (Okawa River, Kokkaido, Japan).

Ramosite—a basic scoria.

Scorlite—a volcanic glass.

Snowflake obsidian—obsidian containing spherulites ranging in size from microscopic to >1 m in diameter.

Zirkelite—an altered basaltic glass.

Economic uses

Volcanic glasses have a variety of economic uses. Pumice is used as lightweight aggregate in concrete and in concrete construction blocks. Volcanic ash is used in the manufacture of pozzolan cements. It is also used as an abrasive in soaps and toothpaste, as well as industrial filtering material and as filler in some paints.

Perlite, after being heat treated, is sometimes used in combination with gypsum in the manufacture of lightweight plaster board; it is also used in horticulture. Basaltic cinder is much used as road metal in regions where it is abundant. Obsidian is still used locally in making jewelry. Unusual varieties, such as snowflake obsidian, that contains small, gray, rosette-like spherulites, and the silky-lustered obsidian of Mexico, are particularly prized by craftsmen.

R. A. BAILEY

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Cross-references: *Lava and lava eruptions*; *Pyroclastic eruptions*; *Tephra*; *Volcanic rocks*.

VOLCANIC NECK

Where ancient volcanoes have been eroded, the underlying conduits through which lavas and other eruptive materials once passed become exposed as volcanic necks or pipes (Fig. 1). Historically, they have also been called "vents," especially in Scotland, but this is no longer appropriate. They are generally subcircular in plan and tend to taper downwards (Fig. 2) although this is not invariably the case.

Necks are commonly assumed to have been drilled by the action of streams of gas and ash ascending ahead of columns of magma. The streams successively fracture, penetrate, and brecciate the surrounding (country) rocks until they themselves become absorbed by stoping and comminution to form part of the streams. At low levels in the structure, or if the process stops before there is eruption at the surface, the streambed material remains intrusive (tuffsite).

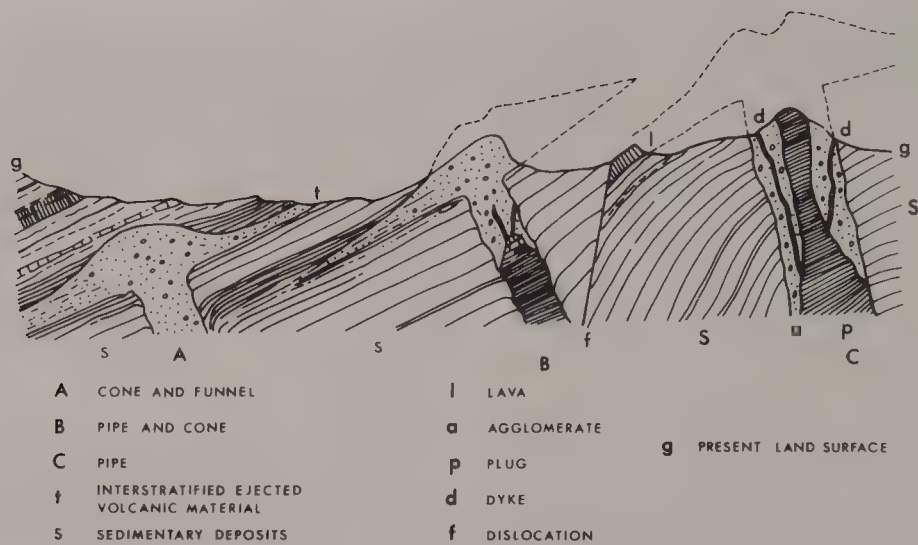


FIGURE 1. Exposure of necks (pipes) when ash-cones and overlying strata are removed by erosion. (After Geikie, 1897.)

In accordance with this concept, tuffisite tends to be well-mixed and homogeneous at the centers of necks where the gas-streaming was more active but to contain high proportions of wall-rock debris, or even to pass laterally into breccias, towards the margins. Streaming, whether by fluidization or turbulent convection (McBirney, 1959), also explains why large xenoliths of country rock within the necks are

displaced by hundreds of meters up or down from their original positions in the walls. The pathway of magma that did reach the surface to become extruded is commonly marked by dykes or cylindrical plugs. These and the pyroclastic neck material that they cut may be fragmented by later eruptions and thus contribute consanguineous blocks to the final composition of the neck. Many pipes are filled

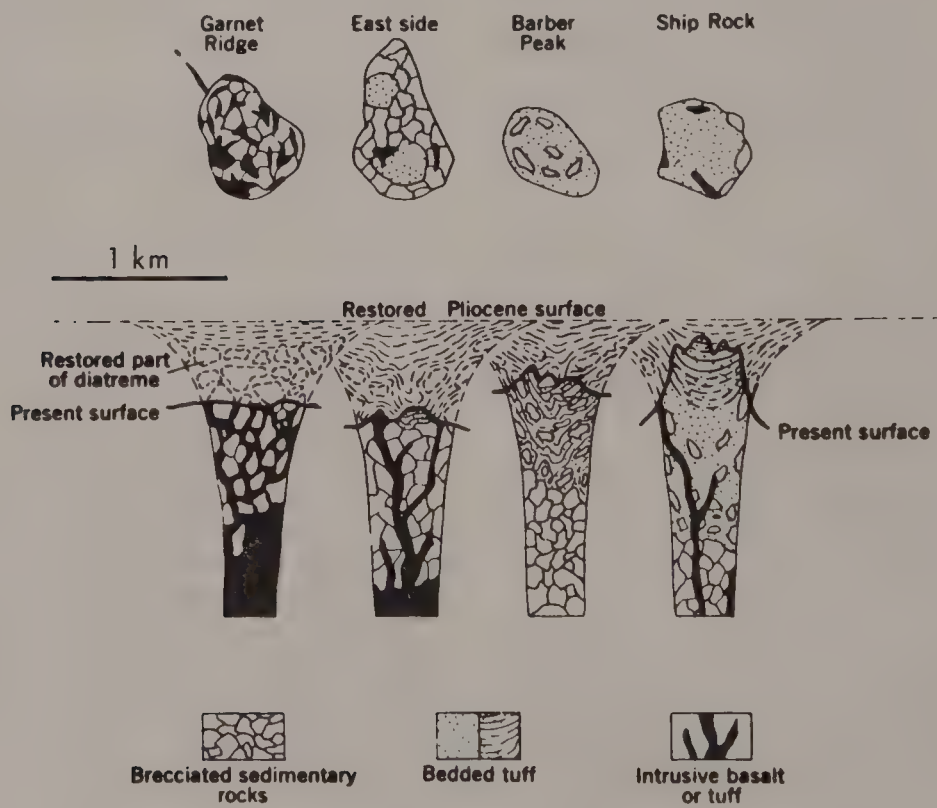


FIGURE 2. Diagrammatic plan views and sections of four necks from the Navajo Reservation, U.S.A., illustrating successive stages of subsidence. (From Shoemaker, 1956.)

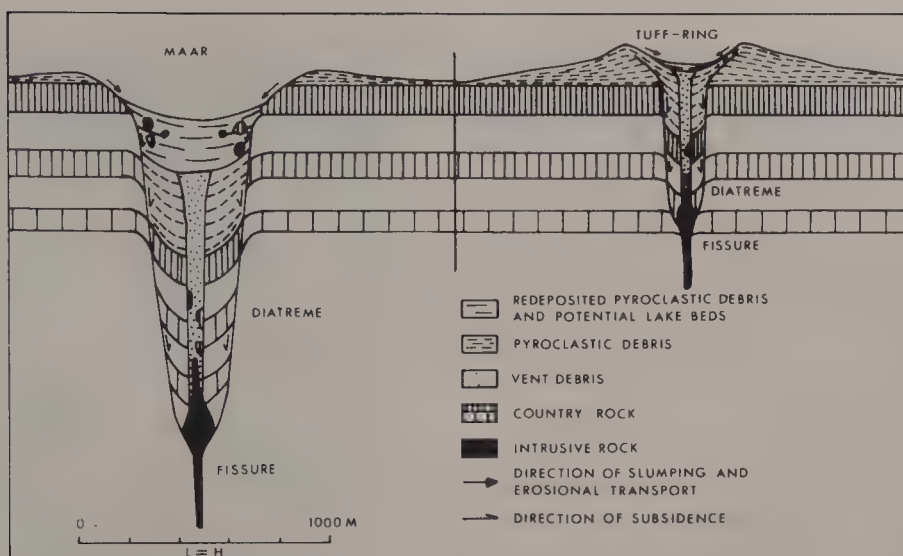


FIGURE 3. Schematic sections through maars and tuff-rings and their feeder necks to illustrate progressive subsidence. (After Lorenz, 1973.)

with bedded tuffs that were demonstrably deposited at the surface and can thus have reached their present position only by later subsidence; descents of more than 500 m have been calculated. The walls of these necks have become redefined as ring faults, the adjacent wall-rocks are dragged downwards against the faults, and the internal bedding is centrocinal, with steep dips adjacent to the faults becoming progressively flatter towards the center. The presence of sediments interbedded with tuffs in some necks suggests alternations of deposition and subsidence over long periods (Shoemaker, 1956).

It now seems clear that necks of this kind represent feeders to maars and ash-rings formed by highly explosive phreatomagmatic activity (Camus et al., 1981) resulting from contact between ascending magma and water, usually at high levels in the crust. Such reaction is characterized by a greater degree of fragmentation than that of strombolian activity, so that a substantial proportion of the resulting tephra (ash) is fine-grained. Bed-forms seen in modern superficial ash-rings closely resemble those of older eroded necks (Leys, 1983) and indicate accumulation by both air-fall and base surge—i.e., lateral blast-plastering (Fisher, 1979)—as well as synsedimentary off-flank slumping into the central maar or ash-ring depressions. Postvolcanic subsidence in the pipes is assumed to be spasmodic, with downward movement following each eruptive pulse so that originally surficial layers descend progressively down the necks (Fig. 3).

It is clear from the evidence of ultrabasic nodules that most necks have deep roots in the upper mantle, probably from the tops of dykes

that extend down to the mantle; this view is reinforced by the many other similarities between such necks and kimberlite pipes (Lorenz, 1975). Closely related dolerite sills in Scotland appear to have been intruded from the upper levels of necks that pierced wet low-density sediments (Francis and Walker, 1987).

E. H. FRANCIS

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Cross-references: *Diatreme*; *Dyke (dike)*; *Emplacement—modes*; *Fluidization*; *Kimberlite*; *Tuffsite*; *Volcanoes and volcanism*.

VOLCANIC ROCKS

Volcanic eruptions produce two main classes of material: lavas and pyroclastic rocks. Lavas are formed by the eruption of molten material (magma) that flows out of a vent or fissure over the surface and solidifies to form compact crystalline, partly glassy, or glassy rock. Pyroclastic rocks are composed of fragmental materials produced during more explosive types of eruption: the finer-grained material makes up volcanic ash and tuff and the coarser material volcanic agglomerate. Lava flows may have a relatively uniform composition, but pyroclastic rocks are more heterogeneous and may contain large amounts of comminuted wall-rocks from the vent as well as fragments of solidified lava.

Lying somewhat between lavas and pyroclastic rocks in their mode of origin are various eruptive products termed pyroclastic flows, including welded tuffs and ignimbrites. They are erupted as fluidized systems in which the liquid phase is dispersed in a gas medium. Such systems flow as discrete bodies termed *nuées ardentes*. These solidify as glassy tuffs in which the fragments represent the remains of originally liquid droplets. In many cases the deposits are hot enough for the glass shards to deform and flow, or weld together, forming compact masses similar to lava. Fragmental structures are often preserved near the base of such units where cooling, in contact with the ground, has been more rapid.

Although the various types of *effusive* product can have almost any composition, there is a strong tendency for the highly viscous, more silicic magma from which acidic rocks crystallize to favor the pyroclastic mode of eruption, while the less silicic, less viscous magma from which basic igneous rocks crystallize commonly erupts as lavas.

Microscopic features of volcanic rocks

Many lavas and welded pyroclastic rocks have a porphyritic texture, in which large crystals (phenocrysts) are set in a fine-grained groundmass (Fig. 1). Most natural magmas do not have

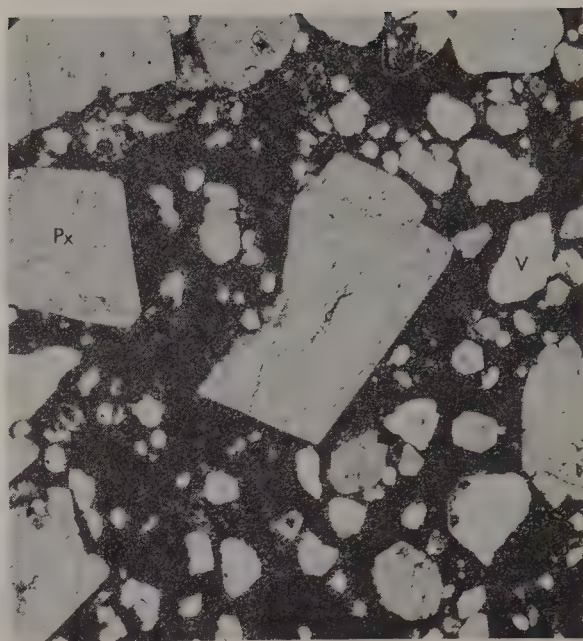


FIGURE 1. Photomicrograph of porphyritic basalt with phenocrysts of pyroxene (Px) in a fine-grained matrix containing vesicles (V); PPL, $\times 25$.

a specific crystallization temperature, but crystallize as they cool through a temperature interval. Most magmas before eruption have temperatures lying within this interval, so that they consist partly of crystals and partly of liquid. Crystals formed during this period of magmatic evolution are termed *intratelluric*, and this includes most phenocrysts observed in volcanic rocks. The process of eruption, with its rapid transport of magma into the cool environment of the Earth's surface, usually results in a great increase in the rate of cooling, and it is at this stage that the fine-grained or glassy groundmass forms.

Rocks devoid of phenocrysts can develop as a result of unusually high magmatic temperatures, but it is more probable that most result from the mechanical separation of intratelluric crystals from their associated liquids (see **Magmatic differentiation**).

The following are other features prominent in volcanic rocks.

Vesicles—voids representing original gas bubbles (vesicular), formed by the process of vesiculation.

Amygdales—vesicles subsequently filled by the deposition of such minerals as chlorite, calcite, chalcedony and zeolites, from percolating vapors and solutions (amygdaloidal).

Flow structure—alignment of mineral grains in response to movements of the lava during the period of final solidification.

Banded structure—produced in a variety of ways but especially characteristic of welded pyroclastic deposits where glass fragments produced by the intense vesiculation and disruption of the liquid phase are intensely flattened by the compaction of the deposit under its own weight.

Apart from these textural features, which provide information about the physical conditions under which particular deposits have been formed, microscopy makes possible the determination of the mineralogical constitution of the rocks. Compared with the plutonic rocks, many volcanic rocks have groundmasses that are too fine-grained to make accurate quantitative mineralogical studies possible. This has important consequences for the classification and nomenclature of these rocks.

Classification and nomenclature

The products of volcanic activity, even of one volcano, are extremely varied in mineralogical and chemical composition. Systems of classification based on chemical analyses are more precise than those based only on microscopic studies. Nevertheless, both types of systems have their advantages; the mineralogical methods, however imperfect, do not require complex analytical instruments.

Important chemical variables include the SiO_2 , Al_2O_3 , Fe_2O_3 , FeO , MgO , CaO , Na_2O , and K_2O content. There are, in addition, important minor constituents such as H_2O ,

TiO_2 , P_2O_5 , and MnO . The major-element variables are not independent and, within reasonable limits, behave in a coherent fashion. Thus, by examining two chemical parameters such as the content of SiO_2 and the total content of Na_2O and K_2O , reasonable predictions of the values of other parameters (e.g., MgO) can be made (Fig. 2). Average analyses on which Fig. 2 is based are given in Table 1. These indicate that the Fe oxides and CaO show a behavior that is qualitatively like that of MgO , while the remaining important constituent, Al_2O_3 , behaves in a rather individual way; it shows much less variation than the other parameters.

These chemical variations are expressed mineralogically with the chemical compositions of the common rock-forming minerals (Table 2), enabling certain generalizations to be made about the mineralogy of various rock types.

1. The more alkali-rich ($\text{Na}_2\text{O} + \text{K}_2\text{O}$ -rich) rocks are likely to contain significant quantities of the alkali feldspars, feldspathoids, or sodic plagioclase. High alkali combined with a high SiO_2 content (e.g., in trachytes and rhyolites) means that alkali feldspars and plagioclase are likely to be important. Conversely, if there is relatively little SiO_2 present (e.g., in nephelinites and phonolites) feldspathoids will be present.
2. If the SiO_2 content $>\text{ca. } 65\%$, free SiO_2 must be present as quartz, tridymite or cristobalite (e.g., in rhyolites).
3. If both SiO_2 and alkalis are low (e.g., in

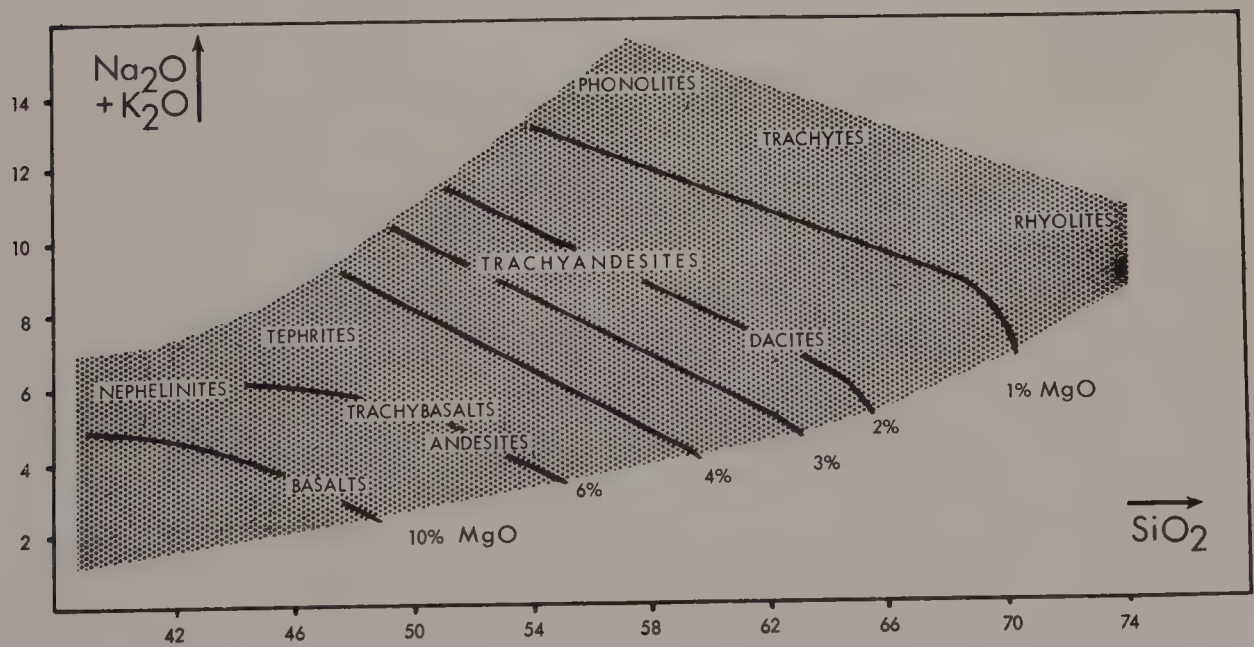


FIGURE 2. $\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. SiO_2 diagram for volcanic rocks. Shaded area is field where the great majority of analyzed rocks plot; names of common rock types are shown; contours give approximate indication of proportions of MgO .

TABLE 1. Average chemical composition of volcanic rock types

	1	2	3	4	5	6	7	8	9	10	11
SiO ₂	74.57	61.95	56.90	63.58	54.20	47.90	45.78	54.02	59.95	44.82	39.07
TiO ₂	0.17	0.73	0.59	0.64	1.31	1.65	2.63	1.18	1.43	2.65	3.86
Al ₂ O ₃	12.58	18.03	20.17	16.67	17.17	11.84	14.64	17.22	19.14	15.42	12.82
Fe ₂ O ₃	1.30	2.33	2.26	2.24	3.48	2.32	3.16	3.83	3.25	4.28	8.75
FeO	1.02	1.51	1.85	3.00	5.49	9.80	8.73	3.98	2.86	6.61	6.39
MnO	0.05	0.13	0.19	0.11	0.15	0.15	0.20	0.12	0.20	0.16	0.26
MgO	0.11	0.63	0.58	2.12	4.36	14.07	9.39	3.87	2.02	7.27	6.14
CaO	0.61	1.89	1.88	5.53	7.92	9.29	10.74	6.76	5.33	10.32	14.20
Na ₂ O	4.13	6.55	8.72	3.98	3.67	1.66	2.63	3.32	6.55	5.30	4.09
K ₂ O	4.73	5.53	5.42	1.40	1.11	0.54	0.95	4.43	4.37	1.26	2.07

Source: after Nockolds (1954).

1 Alkali rhyolite and rhyolite-obsidian

2 Alkali trachyte

3 Phonolite

4 Dacite and dacite-obsidian

5 Andesite

6 Tholeiitic olivine basalt

7 Normal alkali basalt (+ dolerite)

8 Latite (a type of trachyandesite)

9 Nepheline latite

10 Nepheline tephrite

11 Nephelinite

basalt), the ferromagnesian group of minerals are important and the rocks will be rich in MgO and FeO.

Other combinations of minerals are possible, consistent with the major chemical features of a given rock. However, in natural examples certain combinations of the minerals listed in Table 2 are not found, with the following pairs being naturally incompatible—quartz and Mg-rich olivine, quartz and nepheline, quartz and leucite.

The mineralogical variations are summarized in Fig. 3, which illustrates three series whose positions can be seen on Fig. 2 by inspection of the rock-type names. No genetic relationships

are implied by the series chosen; they are three convenient but random sections through the continuum of possible mineral assemblages. Used in conjunction with Fig. 2, they show how the mineralogy and the chemical composition are related to each other.

Problems of identification

Mineralogical charts such as that in Fig. 3 can be used to identify the more common types of igneous rocks provided the specimens are completely or almost completely crystalline; a comprehensive discussion of mineralogical classification is given by Streckeisen (1976). However, volcanic rocks that are wholly or

TABLE 2. Approximate chemical compositions of common minerals of igneous rocks (wt.%)

		SiO ₂	Al ₂ O ₃	MgO + FeO	CaO	Na ₂ O + K ₂ O
Ferromagnesian minerals	{ olivine ^a	40	—	60	—	—
	{ augite ^b	50	3	23	20	—
	{ hornblende ^b	40	10	30	12	tr.
	{ biotite ^b	36	15	30	1	10
Plagioclase series	{ calcic plagioclase ^c	54	29	—	12	5
	{ sodic plagioclase (albite)	68	20	—	—	12
Alkali feldspars	K-feldspar (orthoclase, microcline etc.)	65	18	—	—	17
Feldspathoids	{ nepheline	42	36	—	—	22
	{ leucite	55	23	—	—	22
Silica minerals	quartz etc.	100	—	—	—	—

^a Composition of the common olivine of basic igneous rocks, containing about 25% of Fe₂SiO₄ and 75% of Mg₂SiO₄.

^b In addition to the constituents shown, augite, hornblende, and biotite contain significant proportions of TiO₂ and Fe₂O₃; hornblende and biotite also usually contain about 2% of water.

^c Composition of the common plagioclase of basic igneous rocks, containing about 60% of CaAl₂Si₂O₈ and 40% of NaAlSi₃O₈.

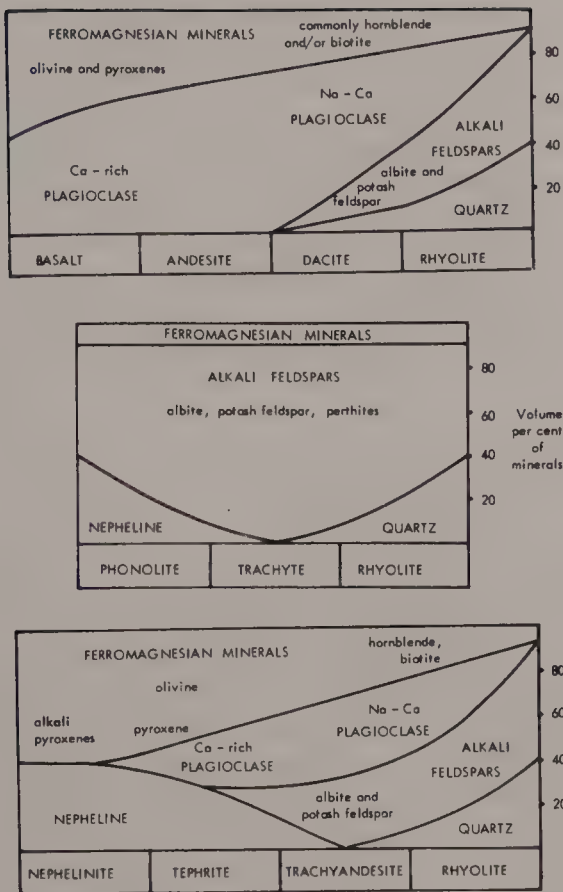


FIGURE 3. Generalized mineralogical compositions of volcanic rock types; positions of the three series can be located on Fig. 1 by inspection of names, but note that no genetic implications are intended by the series chosen.

partly glassy, or that have an irresolvably fine-grained groundmass, present a problem. Here the main clues to identification lie in the nature of the phenocrysts, but the phenocrysts represent only the earlier and higher temperature part of the crystallization history of a rock, and are not necessarily representative of the mineralogy the rock would have achieved if it had cooled slowly. The minerals that generally begin to crystallize at higher temperatures (e.g., olivine, pyroxene and Ca-rich plagioclase) will be over-represented in the phenocrysts, and alkali feldspars, quartz or nepheline, which tend to crystallize late, may not be visible. Identifications based solely on phenocrysts will therefore tend to give a false view of the basicity of the rock—e.g., a partly glassy andesite may appear to be a basalt. For this reason it is desirable to use full chemical analyses for the purposes of identification.

Abundance and distribution

Volcanic rocks have been erupted widely throughout the whole of geologic time; those of

Tertiary or younger age are referred to as being *neovolcanic*. *Basalts* are by far the most abundant type, and are found in all geologic environments. They are characteristic of many currently active volcanoes on mid-ocean ridges, oceanic islands, in island arcs, in cordilleran arcs, and in continental rift valleys. Basalt has also been widely erupted on the Moon and may also be abundant on Mars and Venus.

Andesites appear to be less abundant than basalts, although because of the geologic environment in which they are found they are highly susceptible to erosion. They are found along destructive plate margins, in both island arcs and cordilleran arcs. In the past, these rocks have contributed enormous quantities of detrital sediment to geosynclinal areas.

Rhyolites (and related rocks such as *dacite* and quartz *latite*) are not as voluminous as basalts and andesites, but are of widespread occurrence and form minor components in many igneous environments. They are common along destructive plate margins, and form minor components of volcanic provinces on mid-ocean ridges and ocean islands, and in continental rift valleys.

These three rock types (in the broadest sense) are by far the most abundant volcanic types; others may be regarded as minor and rare, although locally important. *Trachytes* are associated in minor amounts with basalt provinces, particularly with alkali basalts. The more alkali-rich rocks, such as *phonolite*, *nephelinite* (Na-rich), and leucite-bearing rocks (K-rich) are rare overall, but important constituents of ocean island and continental rift valley volcanic provinces. Potassic rocks are also found in other environments, such as the S Italian island arc and the continental plateau areas of southwest U.S.A., Turkey–Iran, and Tibet.

Evolution of volcanic rocks

Many individual volcanoes, or groups of volcanoes, have eruptive products with considerable compositional variation. Investigation of chemical and mineralogical characters indicates that such variations are not random, and often constitute a volcanic series. The general coherence of chemical and mineralogical variables is thus particularly evident in specific cases of associated rocks, or in particular provinces. Figure 4 shows the variation in some chemical parameters for 54 analyzed rocks from Tristan da Cunha in the S Atlantic Ocean (Baker et al., 1964). The rocks concerned vary from low-silica basalts to high-silica trachytes. It is quite common to find that there is no simple time sequence in the way in which the various

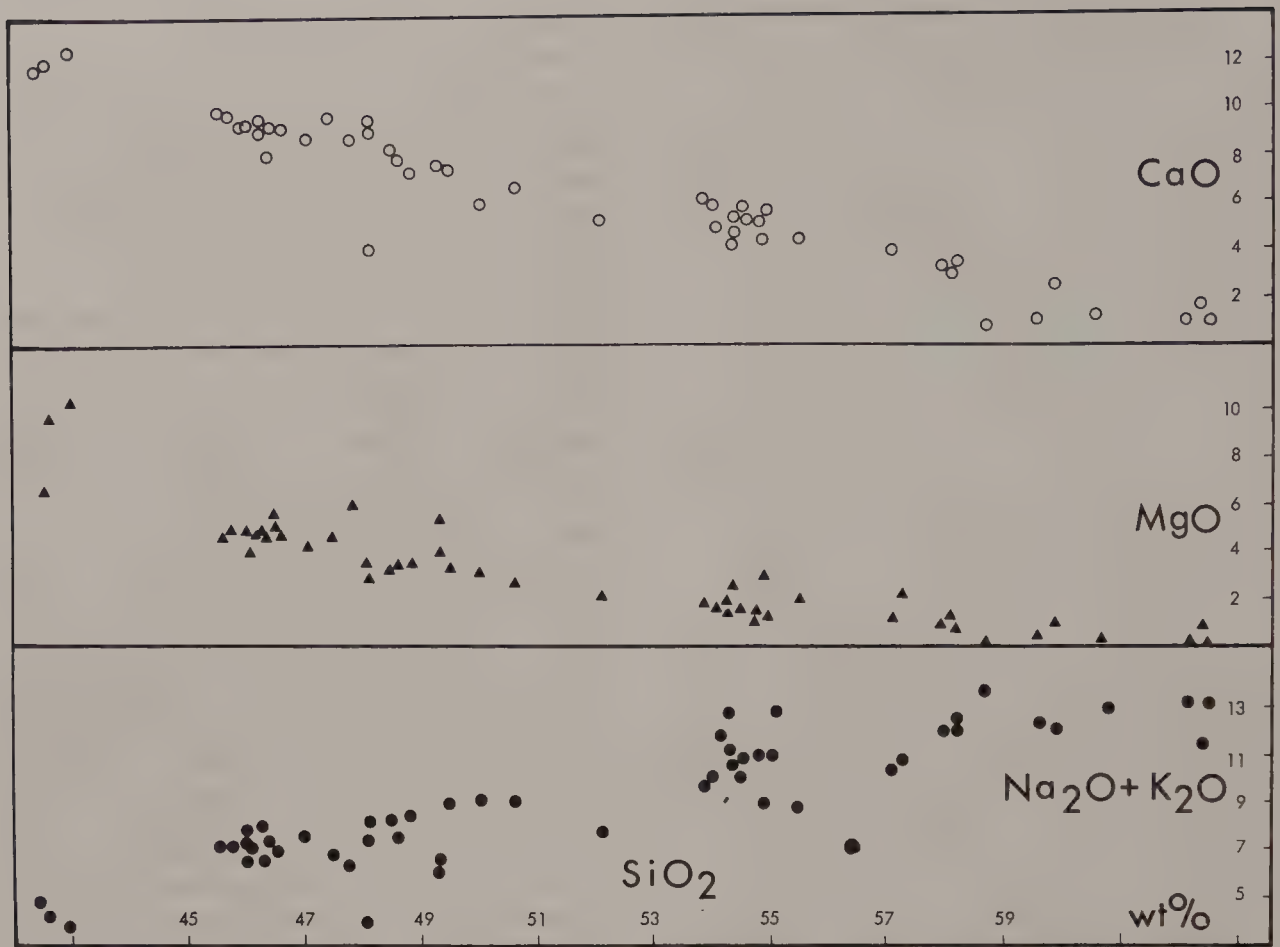


FIGURE 4. Variation of CaO, MgO, and ($\text{Na}_2\text{O} + \text{K}_2\text{O}$) with SiO_2 in analyzed volcanic rocks from Tristan da Cunha, S Atlantic Ocean. (Data from Baker et al., 1964.)

members of the series are erupted. The processes that generate such series are evidently readily reproducible and widespread.

The principal mechanism that has been advocated to explain this type of phenomenon is that of fractional crystallization. It has been demonstrated in many cases that early-formed intratelluric crystals can be separated extremely efficiently from their associated liquids, mainly as a result of sinking under the influence of gravity.

The behavior of basaltic magma illustrates some important principles. The early-formed crystals in basalt commonly include olivine, pyroxene and Ca-rich plagioclase (see Tables 1, 2). The removal of these phases from basaltic liquid will leave a residue enriched in alkalis. What happens to the SiO_2 content is a much more delicately balanced matter, for the mixture of minerals that is removed may have a bulk SiO_2 content not very different from that of the liquid. Hence, depending upon the precise relation of the SiO_2 content of the liquid to the SiO_2 content of the extract, crystallization may result in SiO_2 enrichment or depletion in the residue. If the magma is rather oxidized, so that

magnetite or ilmenite (nonsilicate minerals) appear at an early stage, this will enhance the degree of SiO_2 enrichment considerably.

If the assumption is correct that in many volcanic provinces a magma akin to basalt has a parental role, it is easy to see how the variety of rocks illustrated in Fig. 2 could be produced by crystal fractionation. Alkali enrichment is an almost invariable feature of most stages of fractionation. If this is accompanied by strong SiO_2 enrichment, a series of rocks trending towards rhyolite can be produced, while with lesser degrees of SiO_2 enrichment trachytes or phonolites will be the end products. Concomitant with these effects is the general tendency for Mg, and Ca, and to a lesser extent Fe, to be depleted in the residual liquids.

There is abundant evidence that this general model for the evolution of volcanic liquids is valid. Many studies have shown how the phenocryst sequence in volcanic series are, on the assumption of their removal from the liquid, adequate to explain the variation in the series. Conversely in eroded volcanoes the accumulations of intratelluric crystals (cumulates) that are represented by bodies of gabbro and other

plutonic types, can also be seen to show precisely the required characteristics.

Some volcanic provinces show important exceptions to these rules, notably those that erupt mixed sequences of acidic and basic rocks without the required intermediate members (so-called bimodal series). In continental areas, provinces where there are large volumes of both basalt and rhyolite with almost no intermediate rocks, the rhyolite may have been generated, not by fractionation of a basaltic magma, but by anatexis (i.e., partial melting) of crustal rocks.

Volcanoes as sampling agents of the Earth's interior

Given the nature of magmatic differentiation it should be clear that it is not possible to assume that any liquid emerging from a volcano has been derived directly and without modification from the depths of the Earth. However, the suites of inclusions (xenoliths) they carry up from depth provide important information about the interior of the Earth. Of these, fragments of garnet peridotites, which are found particularly in kimberlite pipes or diatremes, are very significant in establishing an acceptable model for the composition of the upper mantle (see **Pyrolite**).

IUGS nomenclature

The IUGS Subcommittee on the Systematics of Igneous Rocks has made a series of recommendations in order to make the nomenclature of all igneous rocks consistent (Streckeisen, 1976, 1979). What follows is a summary of these recommendations with regard to volcanic rocks.

Names are based on the QAPFM (quartz, alkali feldspar, plagioclase, feldspathoids, mafic) system, and where practical the fields correspond to those for plutonic rocks. The application of this system differs from that applied to plutonic rocks (see **Plutonic Rocks—classification and nomenclature**) in that both modal and normative (CIPW) mineralogy is used (see Streckeisen and Le Maitre (1979) for a comparison of modal QAPFM with CIPW norms). Thus,

$$Q \text{ modal} = Q / (Q + Or + Ab + An)_{\text{normative}},$$

$$\text{and } F \text{ modal} = (Ne + Lc + Kp) / (Ne + Lc + Or + Ab + An)_{\text{normative}},$$

while the A:P modal ratio is expressed as ANOR, where

$$ANOR = An / (Or + An)_{\text{normative}}.$$

The numbered fields on the Q'(F')-ANOR

diagram correspond approximately with those on the QAPF triangle (Fig. 5).

The root names are alkali(-feldspar)rhyolite or liparite, rhyolite or liparite, dacite, quartz-alkali(-feldspar) trachyte, alkali(-feldspar) trachyte, foid-bearing alkali(-feldspar) trachyte, quartz trachyte, trachyte, foid-bearing trachyte, quartz latite, latite, foid-bearing latite, andesite, basalt, phonolite, tephritic phonolite, phonolitic

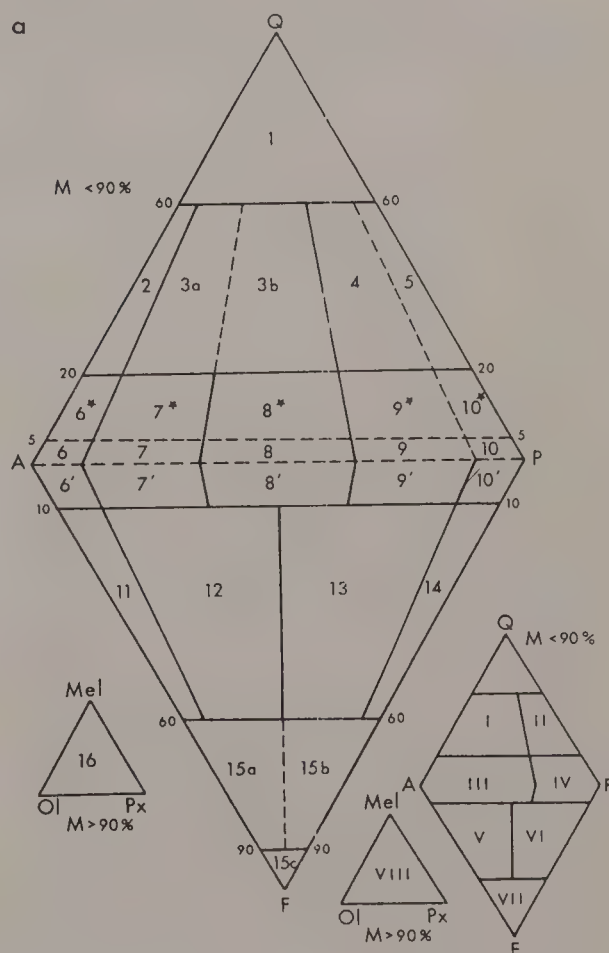


FIGURE 5. Field, modal, and CIPW-normative classification of volcanic rocks after Streckeisen (1979) and Streckeisen and Le Maitre (1979).

(a) QAPFM system with the QAPF triangles for rocks with less than 90% (by volume) total mafic minerals, with inset for mafic-rich ($M > 90\%$) rocks and a field classification.

Root names: 1, none; 2, alkali (-feldspar) rhyolite or liparite; 3a,b, rhyolite (liparite); 4,5, dacite; 6, alkali (-feldspar) trachyte; 7, trachyte; 8, latite; 9,10, andesite and basalt (6*-10* prefixed by "quartz"; 6'-10' prefixed by "foid-bearing"); 11, phonolite; 12, tephritic phonolite; 13, phonolitic tephrite (basanite); 14, tephrite and basanite; 15a, phonolitic foidite; 15b, tephritic foidite; 15c, foidite; 16, ultramafite.

Field names: I, rhyolitoids; II, dacitoids; III, trachytoids; IV, andesitoids and basaltoids; V, phonolitoids; VI, tephritoids; VII, foiditoids; VIII, ultramafitites.

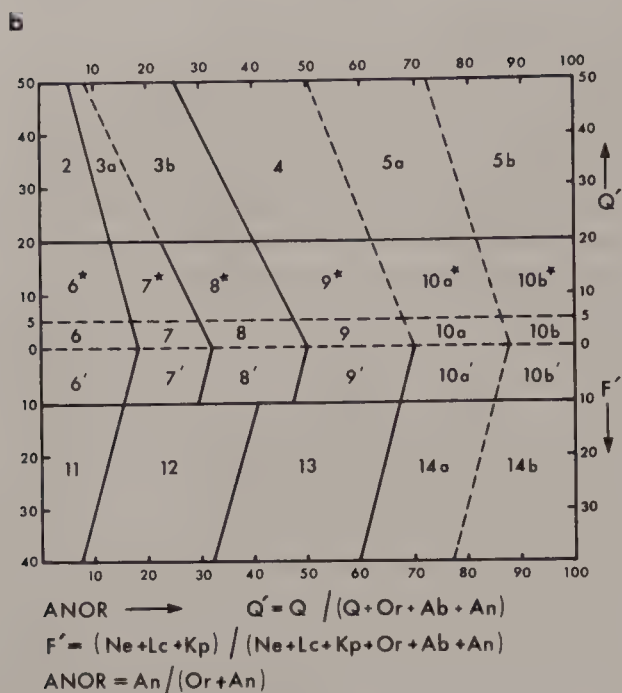


FIGURE 5. (Continued)

(b) Modified QAPF system using CIPW norms, where Q = normative quartz, Or = normative orthoclase ($KAlSi_3O_8$), Ab = normative albite ($NaAlSi_3O_8$), An = normative anorthite ($CaAl_2Si_2O_8$), Ne = normative nepheline $NaAlSi_3O_4$, Lc = normative leucite ($KAlSi_2O_6$), Kp = normative kaliophilite ($KAlSiO_4$)—NB as originally designated and distinct from Ks = kalsilite (K_2SiO_4); root names are the same as in Fig. 5a.

tephrite or basanite, tephrite, basanite, phonolitic foidite, tephritic foidite, foidite, and ultramafitite. (Note that the IUGS recommend that the suffix “-ite” be restricted to volcanic rocks and the suffix “-olite” be restricted to plutonic rocks.)

As in the case with diorite and gabbro, the QAPF system does not discriminate different compositions of plagioclase, so that andesite and basalt occupy the same fields. Feldspar composition and whole-rock SiO_2 content are the only reliable discriminants here.

In all the root names the word “foid” (standing for “feldspathoid”) may be replaced by a mineral name, e.g., leucite basalt, nosean phonolite, nephelinite. Similarly, “ultramafitite” may be given appropriate mineral or defined regional names, e.g., komatiite.

Glass-bearing and glassy rocks should be given a descriptive prefix where the name in the QAPF system is defined using CIPW norms. The IUGS recommendations are 0–20% glass = “glass-bearing,” 20–50% glass = “glass-rich,” 50–80% glass = “glassy,” while 80–100% glass is given special names like obsidian, pitchstone, and tachylite, depending on textures, composition, etc. Porphyritic rocks with a glassy matrix

should be named on the basis of their CIPW norms and given the prefix “hyalo-”, e.g., hyalo-rhyolite, hyalo-dacite.

Metamorphosed volcanic rocks may contain mineral and textural relics, or may retain their original chemistry. In these cases the name should be prefixed by the term “meta-”, e.g., meta-basalt, meta-andesite, instead of the ambiguous names like diabase, epidiorite or leptonite that perpetuate local (or national) traditional usage.

This classification and nomenclature requires a chemical analysis to be applied accurately. Accordingly, a simplified field nomenclature is also recommended. The names used in this are identified by the suffix “-oid” (see Fig. 5).

K. G. COX

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Cross-references: Acid(ic) igneous rocks; Basic igneous rocks; Explosive volcanic eruptions—classification; Flood basalt; Igneous rocks—classification and nomenclature; Igneous rocks—tectonic setting; Intermediate igneous rocks; Kimberlite; Lava and lava eruptions; Mafic and ultramafic inclusions in igneous rocks; Magma; Magmatic differentiation; Mid-ocean ridge and ocean-floor petrology; Ocean-island igneous rocks; Petrochemical calculations; Plutonic rocks—classification and nomenclature; Pyroclite; Volcanoes and volcanism.

VOLCANOES AND VOLCANISM

A volcano is both a vent at the Earth's surface through which magma is extruded as lava, ash, and gases, and the mound of lava and

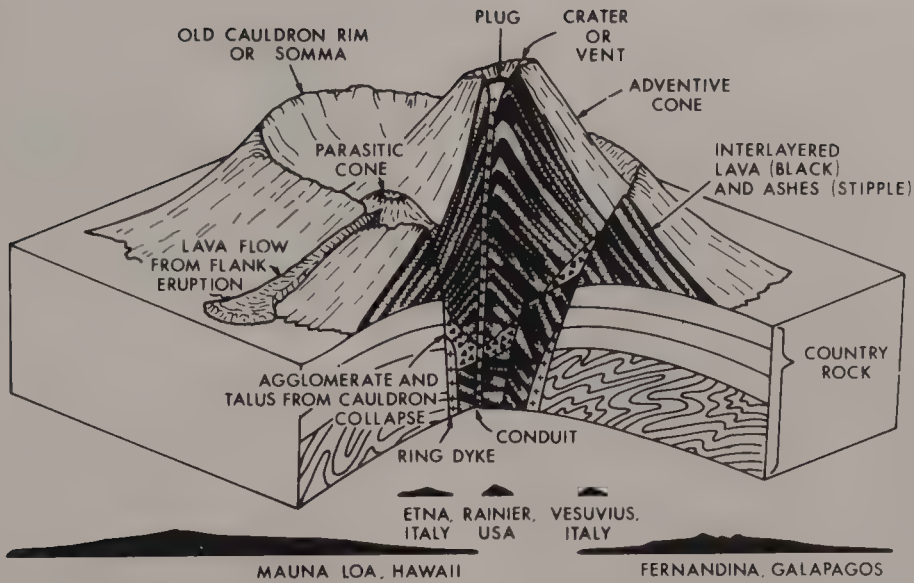


FIGURE 1. Schematic representation of a composite, layered, steep-sided volcano formed when intermediate and acidic magma is extruded and a comparison of shapes of three representative examples with those of shield volcanoes of Hawaii and Galapagos.

ash that accumulates around the vent (Fig. 1). Magma consists of molten rock, usually a silicate melt, with dissolved gases. When eruption occurs the dissolved gases are released and the lava cools and solidifies. On land, the viscosity of the lava determines the extent of lava flow, while the nature of the gas escape determines how explosive the eruption is. Composition (SiO_2 content particularly), temperature of eruption, and the amount of dissolved gas and its solubility in the lava, all contribute to control of lava viscosity (Figs. 2, 3). Acidic magmas have low melting points (ca. 750°C), high viscosity (ca. 10^6 – 10^{12} poise), high gas pressure, but low gas solubility at low

containing pressure. Therefore, acidic volcanism is characterized by explosive eruptions in which large amounts of pyroclastic material are generated. At the same time, such lavas are very viscous and may not be extruded beyond the crater; instead, they may form an endogenous dome or spine. The pyroclastic material may become far-traveled, either as wind-borne plumes or as ground-hugging nuées ardentes or pyroclastic flows. On the other hand, basic magmas have melting points (ca. $1,200^\circ\text{C}$), low viscosity (10^2 – 10^3 poise), low gas pressures, but relatively high gas solubility at low pressures. As a result, on land, basic volcanic activity is characterized by large

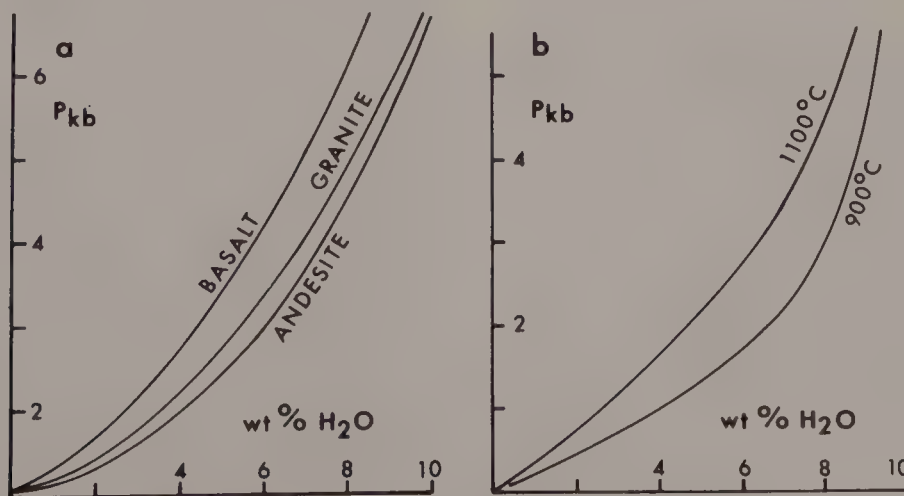


FIGURE 2. Solubility of H_2O in (a) basaltic, granitic, and andesitic magma at $1,100^\circ\text{C}$, and (b) granitic magma at $1,100^\circ$ and 900°C , at varying $P_{\text{H}_2\text{O}}$. (After Williams and McBirney, 1979.)

outpourings of very fluid lava with relatively little pyroclastic material being generated. Intermediate magmas have intermediate properties and consequently their extrusive behavior shares many of the characteristics of acidic and basic activity, without the extremes.

Volcanoes associated with basic magma are frequently enormous edifices with large area:height ratios (e.g., the Hawaiian volcanoes) when they are termed *shield* volcanoes (Figs. 1, 4). Volcanoes associated with acidic magma usually have a small area:height ratio and erupt both lava and pyroclastic material, producing composite, layered, steep-sided cones of *stratovolcanoes*. The same is also the case for volcanoes associated with intermediate magma. This apparently simple classification is complicated by the common occurrence of more than one type of magma being erupted from a particular volcano during its history; this is particularly the case with acidic-intermediate volcanoes, but basaltic shield volcanoes are often capped by alkalic or intermediate stratovolcanoes.

Basalt may also be erupted from fissures rather than central vents, producing so-called *flood* or *plateau* lava fields. Such plateaus may be of enormous extent (see **Flood basalt**), with

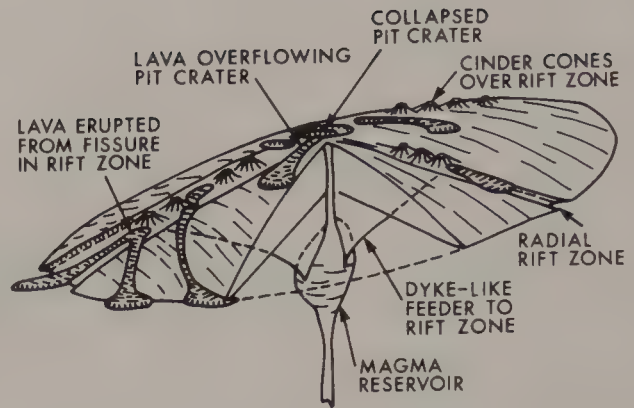


FIGURE 4. Schematic representation of a shield volcano.

the fissures themselves marked at the surface by lines of relatively small cinder cones. Cinder cones themselves are accumulations of pyroclastic material, usually glassy, generated by lava fountains.

Type of eruption

Escher (1933) devised a classification for eruption types that, with modifications, remains in use. It reflects the viscosity of lava and gas pressure in the magma, and indirectly relates explosiveness (the manner of lava degassing) and the amounts and type of pyroclastic material generated (Fig. 5; Table 1).

Hawaii-type eruptions can occur from fissures and central craters, and are characterized by large volumes of highly fluid lava (usually basalts or mafic feldspathoidal alkalic rocks). Lava fountains produce cinder cones and the containment of lava in a crater may produce long-lived lava-lakes, such as those of Kilauea (Hawaii), Samoa (SW Pacific), Nyiragongo (Zaire), and Erebus (Antarctica). Pyroclastic rocks tend to be autoclastic or vitreous, and comagmatic with the lavas.

Stromboli-type eruptions are characterized by mildly explosive activity over a prolonged period from summit craters. The activity is rhythmic over a period of hours or weeks, and produces mainly cinders or bombs in large quantities. Stromboli in the Lipari Islands is the type example; others include Sakurajima (Japan) and Irazu (Costa Rica). Strombolian activity can last for many years, and there is a relationship between the length of time between outbursts and their violence.

Vulcano-type eruptions vary between weakly and strongly explosive, and like strombolian eruptions produce mainly pyroclastic material. They typically produce large, vertical eruption clouds during short outbreaks separated by long periods of quiescence.

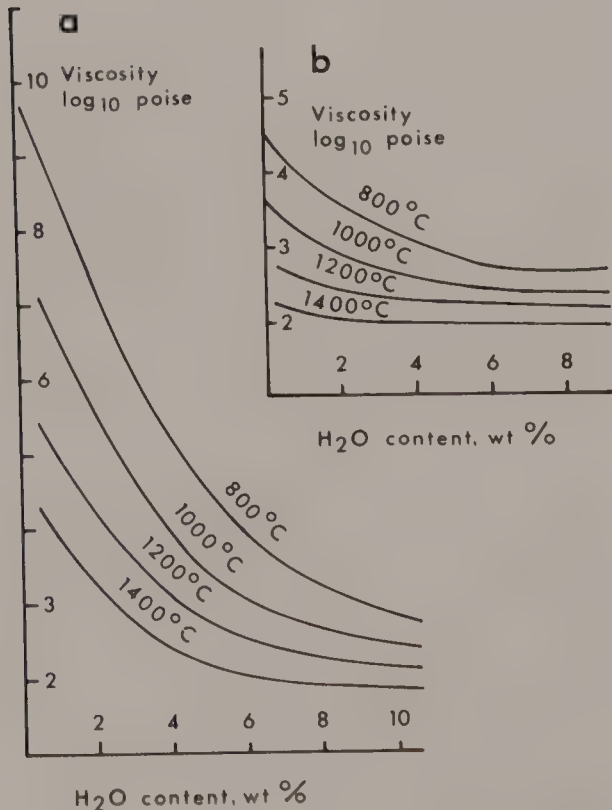


FIGURE 3. The effect of H₂O content on magma viscosity at various temperatures, for (a) granitic and (b) basaltic liquids. (After Williams and McBirney, 1979.)

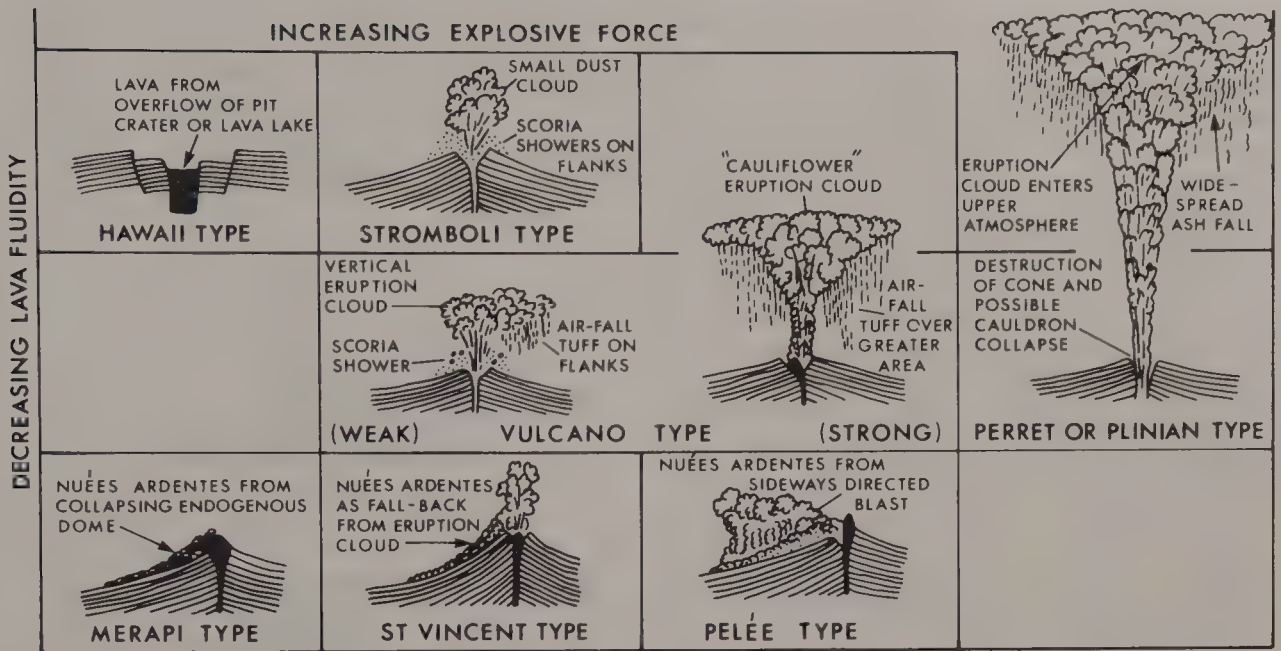


FIGURE 5. Escher's (1933) classification of eruption types by degree of explosiveness and lava fluidity.

Plinian- or Perret-type eruptions represent the extreme of a spectrum that includes strombolian and vulcanian types. They are extremely violent, generating enormous quantities of pyroclastic material and a vertical eruption cloud that may rise into the stratosphere. Unlike the previous two types of activity, which result in a growing volcanic edifice, blasts during plinian-type eruptions usually accompany the destruction of the edifice, often during cauldron collapse or caldera formation, e.g., Katmai, Alaska (1912), Krakatoa, Indonesia (1883), Tambora, Indonesia (1815), El Chichon, Mexico (1982), and the type example, Vesuvius, Italy (AD 79) (see **Calderas and volcano-tectonic depressions**).

Merapi-, St. Vincent- and Pelée-type eruptions represent progressively more explosive examples of an eruption type associated with highly viscous lavas, and all include nuées ardentes among their characteristic features. Typically, the lava erupted is so viscous that it does not flow beyond the crater, but is forced from the vent as an endogenous dome or spine-like plug. Explosive outbursts usually follow some event that acts to release pressure on the rising magma, precipitating explosive degassing. Slump collapse of an endogenous dome, or part of the flanks of the volcano, may result in a huge, sideways-directed blast, in which pulverized rock from the volcano and incandescent froth from the new lava are carried horizontally as a fluidized mass traveling at speeds in excess of 70 km/hr, e.g., Merapi, Indonesia (1930); Mt. St. Helens, U.S.A. (1980); Bezymianny, U.S.S.R. (1956). This

represents peléan activity, effectively a sideways-directed plinian eruption.

The St. Vincent-type represents a pressure release confined within a crater, possibly due to endogenous dome collapse, in which the eruption cloud rises vertically at first, but considerable material falls back onto the volcano flanks, generating glowing avalanches down the slopes. In a Merapi-type eruption, nuées ardentes and glowing avalanches issue from the collapsing snout of a viscous lava flow, or from the flanks of a collapsing endogenous dome, effectively representing the violent degassing of lava already extruded.

Phreatic activity

This is the interaction of hot rock with water under pressures low enough for steam to form, resulting in explosive activity. This occurs on several scales, related to different eruptive environments.

Submarine eruptions that occur at water depths at which steam can form produce distinctive eruption products. Whatever the nature of the extruding lava, the result is explosive decrepitation of quenched lava, producing glass-shards (hyaloclasts) and brecciated rock. The resulting deposit is termed *aquagene tuff*, *hyaloclastic tuff*, or *hyaloclastic breccia*. Similar activity accompanies extrusion of lava over or into wet, unconsolidated sediment; the resulting mixture of lava and comminuted lava with sediment is termed *peperite*.

Strictly, phreatic activity refers to explosive,

steam-driven eruptions whose products are not primary magmatic material, but the results of recent research into pillow-lava formation and volcanic activity in the shallow marine environment have led to a broadening of the concept covered by the term phreatomagmatic. This includes the formation of pillow lavas, hyaloclastites, and pepperites, and characterizes an eruption type not considered by Escher (1933)—*Surtsey-type eruptions*. Surtseyan activity accompanies the growth of a submarine volcano as it approaches sea-level and emerges above the high-water level, the island of Surtsey, south of Iceland being the type example.

Surtseyan activity can accompany the eruption of lava of any composition, and is explosive owing to the flooding of an active crater by water. Steam explosions generate large quantities of pyroclastic material, mainly hyaloclastites that subsequently alter to palagonite. The eruption site is rapidly surrounded by a rampart of such material (palagonite tuff-ring), which is continuously being attacked and breached by wave action. Such an island is a temporary feature, being rapidly denuded once the eruption ceases unless it is capped by lava. Massive lava can only be extruded once the tuff-ring becomes an effective dam, preventing phreatic brecciation of lava.

One of the biggest and best documented phreatic eruptions of modern times was that of the Tarawera fissure in New Zealand (1886). A fissure 15 km long was the site of a great variety of activity, much of which was phreatic and very little resulted in fresh lava being erupted beyond the fissure itself, and this was largely restricted to rhyolite domes. Some 5 km of the fissure ran beneath the basin now occupied by Lake Rotomahana, and here two small lakes drained into the fissure. Massive phreatic explosions outside this basin produced extensive deposits of pyroclastic material, both decrepitated fresh lava and brecciated lavas from earlier eruptions; but within the basin the eruption expelled approximately 1 km³ of mud.

Subglacial eruptions

This type of eruption is best documented on Iceland; distinctive deposits and land-forms are produced. When volcanoes erupt beneath an ice-cap, large volumes of meltwater are generated (*jokulhlaup* in Icelandic), so that many features characteristic of subaqueous and phreatic activity result. If the eruption fails to break through the ice-cap, the resulting edifice is a pile of pillow lava and hyaloclastites; when the eruption breaks out, the pile may be capped

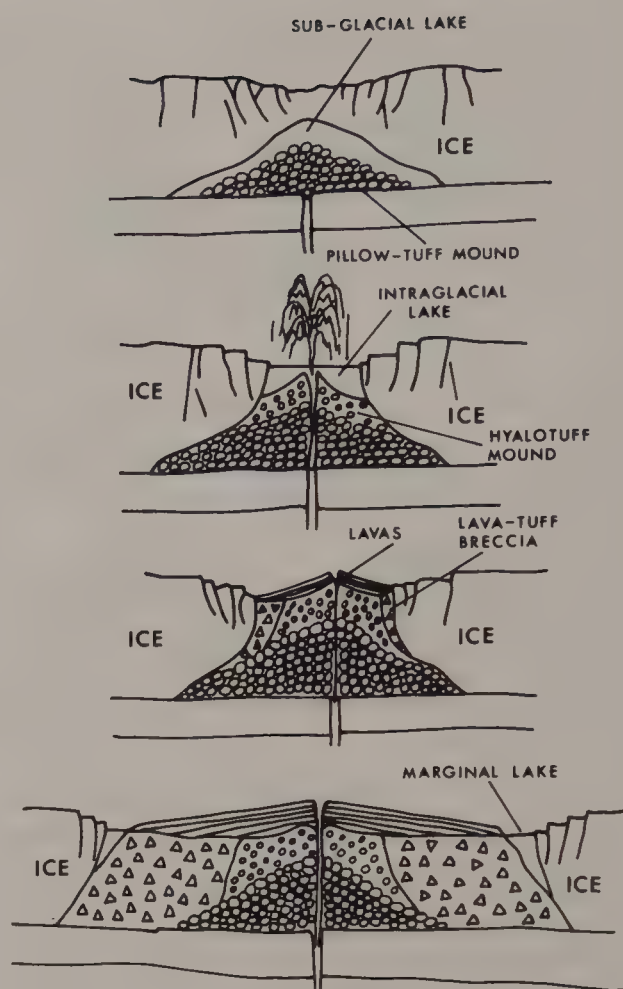


FIGURE 6. Stages in the formation of an Icelandic table mountain during subglacial eruptions.

by massive lavas. Subsequent melting of the ice-cap leaves behind a steep-sided *pillow-tuff mound* or ridge (*moberg* in Iceland, *tuya* in British Columbia, Canada), or table mountains capped by lava (Fig. 6).

Non-eruptive volcanic activity

As the eruptive activity dies down at a volcanic center (*decadent stages of volcanism*), the distinction between phreatic activity and hydrothermal activity becomes blurred. Water circulation may be an important mechanism in the cooling of thick pile of lava or pyroclastic flows. Geysers, hot springs, geothermal fields, hot ground, solfataras and fumaroles are frequently located within calderas, volcano-tectonic depressions, or lava and pyroclastic fields. Such circulating waters are often rich in dissolved minerals, especially S and Si, and may be a prime agent in zeolite facies metamorphism. This is referred to as the solfataric stage of volcanic activity.

TABLE 1. Magma composition and eruptive activity

<i>Acidic</i>	Very volatile-rich; low gas retention	Low eruption temperature; high viscosity	Explosive to very explosive; high % fragmental products
<i>Intermediate</i>	Volatile rich; variable gas retention	Medium range eruption temperature; medium range viscosity	Explosive to effusive; varying % fragmental products
<i>Basic</i>	Low volatile content; high gas retention	High eruption temperature; low viscosity	Effusive; low % fragmental products
<i>Alkalic</i>	High to low volatile content; high to low gas retention	High eruption temperature; variable viscosity	Very explosive to effusive; variable % fragmental products

Changes in volcanic activity with time

The nature of volcanic activity at a particular site may vary with time; for instance, as a volcanic edifice rises from the sea floor and emerges above sea level, or as a magma suite evolves. Likewise the style of eruption may vary with time, building up a composite volcano. Many large volcanic edifices have gone through a growth stage, characterized by hawaiian,

strombolian, vulcanian, or peléan activity, that is terminated by a paroxymal plinian eruption accompanied by cauldron collapse. After a period of quiescence, a new constructive phase begins with secondary (or adventive) cones being built within the caldera. Such a composite edifice is sometimes called a somman volcano, where the caldera rim is termed a somma (Fig. 1). The type example is Monte Somma with its adventive cone (Vesuvius) near Naples, Italy.

TABLE 2. Distribution of active and recently active volcanoes

Area	Active ^a	Holocene	Solf., fum., hot spr.
Europe, Asia Minor,			
Iran, Caucasus	12	20	13
Arabian Pen., Levant	7	15	4
Africa, Red Sea	24	31	36
Indian Ocean	6	8	1
Australasia, Melanesia	59	14	41
Indonesia, Andaman Is.	80	5	44
Philippines, Taiwan,			
Ryukyu Is., Micronesia	54	15	19
Japan	34	10	14
Kurile Is., Kamchatka	54	42	13
Mainland Asia ^b	4	14	1
Aleutian Is., Alaska	44	27	6
W Canada, W U.S.A.	9	64	6
Hawaiian Is., Galapagos,			
Polynesia	19	11	4
Mexico	14	13	5
Central America	32	32	13
Colombia, Ecuador, Peru	21	6	6
Bolivia, Chile, Argentina,			
Juan Fernandez Is.	35	38	23
W Indies	7	5	5
Iceland, Jan Mayen	29	37	0
Atlantic Ocean (S of Iceland)	24	4	2
Antarctica	11	15	8
Arctic Ocean	2	^c	0
Total	581		

^a Active is defined as having a documented eruption history; Holocene indicates evidence of eruption during the last 10,000 years; Solf., fum., hot spr. indicates a volcanic center with no evidence of eruption in the last 10,000 years, but a site of solfataric, fumarolic, or hot spring activity.

^b Mainland Asia excludes Asia Minor, Arabian Peninsula, Levant, Iran and Kamchatka.

^c No data

Source: data from Simkin et al. (1981).

Other examples include Crater Lake (Mt. Mazama), U.S.A., Taal, Philippines, and Masaya, Nicaragua.

Distribution of volcanoes

Most volcanic activity occurs unseen on the crest of the mid-ocean spreading ridge system; pillow lavas and massive basalt flows are extruded there into the axial rift zone. Elsewhere, volcanism occurs along destructive plate margins above subduction zones in island arcs or cordilleran volcanic chains. The volcanism here is more diverse—the products include tholeiitic basalt, andesite, dacite, and rhyolite of the calc-alkaline suite. Large quantities of pyroclastic material are generated from explosive activity centered on stratovolcanoes and composite shield–stratovolcanic piles. In the cordilleran belts, however, volcanism is not evenly spread above the active subduction zones, and in the Andes, for instance, several volcanic “quiet zones” exist. The majority of active subaerial volcanoes are located on these destructive margins, particularly in the “Ring of Fire” in the circum-Pacific region (Table 2; see **Circum-Pacific plutonic and volcanic activity**).

Away from these plate margins, active volcanoes are associated with rifting, e.g., the intracontinental rift valleys like those of E Africa, and the uplifted continental plateaus like Tibet and southwest U.S.A. In the intracontinental rift valleys, basalts are erupted alongside a host of alkalic, undersaturated rocks including phonolite, trachyte, and feldspathoidal lavas, like those erupted from the twin shield volcanoes Nyiragongo and Nyamalagira in Zaire. Some unusual varieties of volcanic activity include the Na-carbonate lavas of Oldoinyo Lengai, Tanzania.

Hot spots occur on both oceanic and continental crust, producing isolated volcanic centers, whose locus may have shifted with time. This activity is often basaltic and intense, including Iceland and Hawaii, the two most volcanically active areas on Earth.

Some volcanic phenomena known from the geological record do not appear to have been witnessed by man, and as a result their exact nature is rather obscure. These are (1) the lava shields of Iceland and Hawaii which, although currently active, do not compare in scale with the enormous flood-basalt plateaus of the previously existing supercontinent Gondwanaland or the Columbia River, northwest U.S.A.; (2) the huge sheets of ignimbrite thought to have erupted from fissures in New Zealand, Chile, and the Yellowstone area of Wyoming, U.S.A., which have no modern equivalent,

although both the Taupo depression and the Yellowstone caldera are still active; (3) the eruption of kimberlite diatremes, probably violently explosive, although the pipes in the Igwisi Hills, Tanzania may be less than 1 Ma old; and (4) the eruption of ultramafic komatiite, which appears to be restricted to the older part of the geologic record, and for which temperatures in excess of 1,600°C and viscosities of below 10^2 poise have been deduced.

Volcanic hazards

The impact of volcanic phenomena on man depends not only on the form of the activity but on its proximity to densely populated areas. Lava flows present few problems to life, their movement being predictably controlled by topographic features. Their potential for damaging crops and property, either by inundation or by causing fires is considerable. The greatest dangers are posed by the more violent manifestations of peléan-, merapian-, and plinian-type eruptions, and some phreatic phenomena. Nuées ardentes and mud-slides (lahars) generated by torrential rain, the expulsion of crater lakes, or melting ice caps have all caused extensive loss of life and damage to property.

Two manifestations of volcanic activity can have more widespread effects, one dramatic, the other more subtle. Tsunamis or seismic ocean waves have been associated with major eruptions on volcanic islands, particularly with the 1883 eruption of Krakatoa in Indonesia, when some 30,000 deaths were attributed to the sudden inundation of low-lying coastal areas of Sumatra and Java. Plinian eruption may inject enormous quantities of dust and sulfurous aerosol into the upper atmosphere, where they may remain for months or even years. Such a dust or aerosol cloud may have climatic effects. The plinian eruptions of Tambora, Indonesia (1815), Gunung Agung, Indonesia (1963), and El Chichon, Mexico (1982) were all apparently associated with worldwide climatic effects.

Other terms

Littoral cone—a small volcanic cone formed when a lava flow enters a body of water.

Mofette—CO₂ exhalation, and the small opening from which the gas is emitted in an area of late-stage volcanic activity.

Puy—a small remnant cone, particularly in the Auvergne district, France.

Tumescence—swelling of a volcanic edifice resulting from magma accumulation in the reservoir; sometimes followed by eruption.

Ultravulcanian—type of eruption characterized by violent, gaseous explosions of lithic dust and blocks with little, if any, incandescent scoria, as at Krakatoa in 1883.

Volcanello—a small volcano such as an active cone within a central crater of a volcano (Mt. Nuevo in Vesuvius, Italy), or a spatter cone.

ADRIAN F. PARK

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- Cross-references: *Calderas and volcano-tectonic depressions; Circum-Pacific plutonic and volcanic activity; Endogenous dome; Explosive volcanic eruptions—classification; Flood basalt; Fumarole; Geyser; Igneous rocks—tectonic setting; Island arcs—petrology; Lava and lava eruptions; Magma; Mid-ocean ridge and ocean-floor petrology; Nuée ardente; Ocean-island igneous rocks; Palaeo-volcanology; Pyroclastic eruptions; Pyroclastic flows; Rift valley volcanism; Solfatara; Volcanic glass; Volcanic rocks; Zeolite facies.*

W

WATER IN ROCK MELTS

An active volcano is one of nature's greatest spectacles and many of the more violent events are due to exhaled gases. The analyses of these exhalations show that water is usually the major component. Its role in modifying the physical properties of magma, in inducing the crystallization of hydrous phases, and serving as a reservoir of oxygen when dissolved in a magma, is much less spectacular than the role in volcanism, but petrogenetically speaking much more significant.

Water saturation level of magmas

There seems to be no concerted opinion among geologists on a reasonable figure to assign to the water content of magmas. Estimates range from essentially zero to the saturation value at some considered pressure.

Evidence from active volcanoes. Several observers have attempted to estimate the amount of gas given off by a volcano in a single eruption (Macdonald, 1963). These estimates have been based on the size of the volcanic opening and the velocity of the outward-rushing gas as evidenced from the height to which dust is carried. If the amount of lava involved in the same eruption can also be estimated, a very rough figure for the amount of water in the magma can be calculated. One serious objection to this method is that no account is taken of how much of the observed water vapor has never been in solution in the lava but is merely heated meteoric or seawater (Friedman et al., 1974). Many geologists consider that all the water vapor involved in volcanism is of surface or near-surface origin. One of the earliest successful attempts to collect gases from an active volcano was made by T. A. Jaggar in 1919. At that time, the Kilauea crater, Hawaii, had a lake of molten lava and the conditions for the collection of gases directly from the lava were excellent and have been unsurpassed since.

The average water content of 14 samples taken by Jaggar (reported by Shepherd, 1938) is 70-53 vol.%. Heald et al. (1963) took samples from the 1959-60 eruption of Kilauea. This time no lava lake was present and the samples were

taken from a large cinder cone. The water content of these samples was 97-99 mol.%; subsequent calculations on the equilibrium of the gaseous species making up the other 1-3% of the vapors indicated that they could not have been in equilibrium at the observed temperature of the magma. However, if only 50% of the gas was water vapor, the observed and calculated conditions of equilibrium would agree closely.

The conclusion that can be drawn from this work is that volcanic gases are easily contaminated with water when in a near-surface environment and any calculation of total gaseous discharge from a volcano may be much in error.

Water content of rocks. Water in rocks is usually in the structure of hydrous minerals, or in glass, or as molecular water in fluid inclusions within mineral grains; there may also be some adsorbed water between mineral grains. If the rock is free from alteration and weathering effects (i.e., if it has not gained water after consolidation), the water content gives a minimum value for the water content of the parent magma, since it is not known how much water evolved during crystallization.

Intrusive rock magmas cooled at depth in the Earth's crust crystallize slowly under considerable rock pressure. The slowness of the cooling process precludes the formation of glass and if any water is retained by the rock it will be mostly in the structure of hydrous minerals plus minor amounts of water in inclusions. The two most common hydrous mineral groups of igneous rocks are amphiboles and micas, whose water contents average ca. 1.7 and 3.5 wt.%, respectively, so no matter what the starting water content of the magma, the water content of the rock is limited. It is not surprising, therefore, to find the average water contents of granite and gabbro, as computed by R. A. Daly to be 0.84 and 1.45 wt.%, respectively. That much water is lost from magmas during crystallization is evidenced from hydrothermal alteration in pre-existing rocks around the intruded igneous rocks, especially granitic rocks.

The relatively fast cooling of magma extruded on the Earth's surface improves the chances of retaining most of the magmatic water. Even in the case of the very viscous rhyolites, however,

the formation of vesiculated lava proves that in many cases the lava is not quenched quickly enough to keep all the water in solution. Basaltic magma is so fluid that it is never quenched quickly enough to stop most of the melt crystallizing. Therefore, the water content of extruded rocks is merely the amount of water that was retained during the quench, and is probably more dependent on the rate of cooling than on the initial water content of the magma.

Since the water vapor pressure (P_{H_2O}) in the atmosphere is very low, a magma on or near the surface will tend to lose essentially all its dissolved water. It is surprising, therefore, that Daly's computed averages show the average water content for rhyolites and basalts to be 1.47 and 1.62 wt.%, respectively. Either the quenching of the lavas is fast enough to retain that much water or the analyses considered by Daly included a number from altered lavas. Modern analyses of lavas show less water; for example, the average water content of 25 Hawaiian lavas is 0.23%.

Other methods. Several other methods involving models of a primordial Earth have been used to calculate the amount of water that might be expected in magmas. These calculations usually assume that all the water now at the surface was originally in the molten skin of the Earth. The assumptions involved here are very speculative and the results can only be expected to be grossly applicable to modern magmas. Gilluly (1937) made such a calculation and concluded that the average percentage of water was about 7. Ingerson (1950) assuming an entirely different model, calculated the water content of the primitive magma to be about 3%.

Experimental determinations of water content of rock melts. The foregoing brief survey of the application of natural observations to the problem of the amount of water in rock melts leads to the conclusion that no satisfactory solution can come from such methods. At present, experimental results cannot offer an answer to the question of the amount of water in natural rock melts, but they can at least provide data on the effect of pressure, temperature, and composition on the saturation or maximum possible water content. The three curves in Fig. 1 labeled B, G, and A show the maximum amount of water that can be dissolved at 1,100°C in a basaltic, a granitic, and andesitic melt, respectively, as a function of pressure. The effect of temperature can be estimated by comparing the granite curve determined at 1,100°C with the pegmatite curve (curve P), which was determined at about 700°C. This figure shows that (1) the water solubility in rock melts increases with pressure with the curves showing no signs of flattening off in the pressure

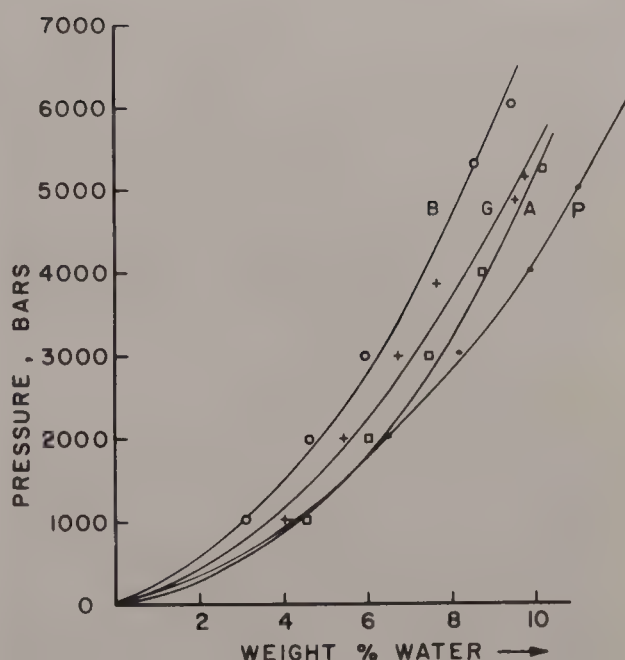


FIGURE 1. The solubility of water in four rock melts as a function of P_{H_2O} . B (open circles), a Columbia River basalt; G (crosses), Westerly granite; A (squares), a Mount Hood andesite; and P (filled circles), Harding pegmatite. B, G, and A were determined at 1,100°C; points on P were determined just above the liquidus temperature and all are within $\pm 25^\circ\text{C}$. (A, B, and G from Hamilton and Anderson, 1967; P from Burnham and Jahns, 1962.)

range considered; (2) the effect of melt composition on solubility is small, the difference between the basaltic and granitic melts at 6,000 bars being only 1%; and (3) the effect of raising the temperature is to decrease the solubility.

The determination of water solubility as a function of pressure, temperature, and composition can also help an understanding of both the mechanism of solubility and the structure of the melt. For example, Oxtoby and Hamilton (1978) have shown that a graph of water solubility against $(\text{pressure})^{1/2}$ is a straight line that predicts that water is held as (OH) and not as H_2O in the melt. Further consideration of their data suggests that a tridymite-type structure is common and that Al tends to move from the tetrahedral to octahedral position with the addition of water.

Comparative phase-equilibria studies. Maaløe and Wyllie (1975) have deduced that the parent magma to a biotite granite from the Bohus batholith of Norway-Sweden had a maximum of 1.2 wt.% water. Their conclusion is based on the fact that the order of appearance of the minerals deduced from the textures of the rock (magnetite, plagioclase, microcline, quartz, and finally biotite) is only reproduced in

experiments at 2 kb when the amount of water in the starting material is $<1.2\%$.

Another estimate of the water content of a granite magma based on phase-equilibria data has been made by Day and Fenn (1982). In this case the conclusion, which depended on results from a synthetic granite system (James and Hamilton, 1969), was that the parent magma to the Scituate granite, Rhode Island, U.S.A. probably had <4.7 wt.% water.

Effect of water on rock melts

Effect of water on melt viscosity. The viscosity of magmas is a very important physical property (the unit of measurements is the poise). Various dry melts at 1 atm. and at a temperature just above the liquidus temperature have the following viscosities: rhyolite, 10^{11} ; andesite 10^7 ; basalt 10^3 (Murase and McBirney, 1973). The cause of the very high viscosity of rhyolite melt is its very polymerized structure, resulting from an extensively linked Al, Si-O framework. The other two melts are much poorer in Si and are thus very much more fluid. Viscosity of the melt is probably the most important property controlling the mode of occurrence of the corresponding rock; i.e., the most common occurrence of rocks of basaltic composition is as lava flows, with a much lower proportion occurring as intrusive gabbros, whereas the very viscous rhyolite liquid rarely occurs as flows but often congeals at depth to make granite plutons. The velocity of ascent of a magma is inversely proportional to its viscosity.

The addition of water to a rock melt breaks up the Al, Si-O structure and thus reduces the viscosity. Accordingly the effect of a given amount of water on the viscosity of rhyolite liquid is predictably much greater than on a basalt liquid, and this has been borne out by experiments. The determination of the viscosity of hydrated melts must be done at a pressure to keep the water in solution: the experiment is a difficult one and not surprisingly only a few data are available. Figure 2 shows results obtained for rhyolite melts: 4 wt.% water reduces the viscosity from ca. 10^{11} to ca. 10^6 poise, an enormous decrease. Reference to Fig. 1 shows that a minimum of 1 kb of pressure (≈ 3.5 km depth in the Earth's crust) is needed to keep 4% water in solution. In a series of high-pressure experiments Kushi et al. (1976) have measured the viscosity of a hydrous andesite melt. The rate of fall of a dense sphere (olivine, or pyroxene or platinum) was measured and then Stoke's law was used to calculate the viscosity; the density of the melt was taken to be that of the glass, formed by quenching the melt. It was

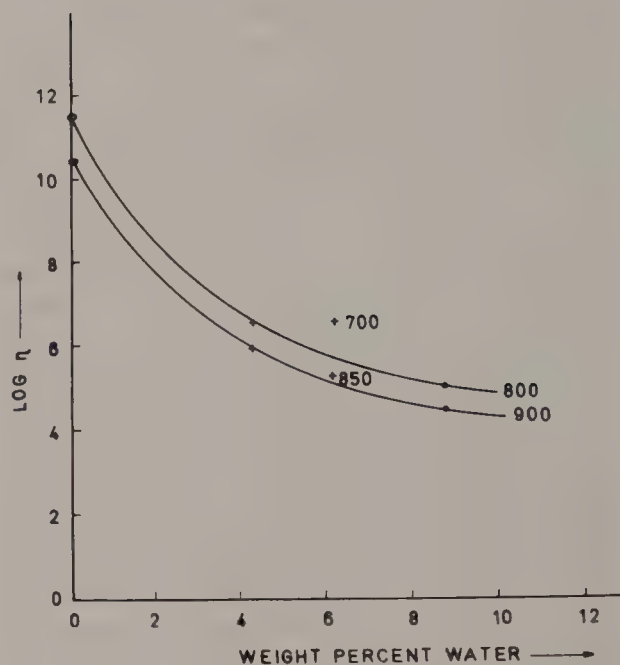


FIGURE 2. Viscosity (poises) of melts of rhyolitic composition as a function of water content; open circles, crosses, and filled circles plotted from data in the *Handbook of Physical Constants* (1942), Shaw (1963), and Burnham (1963), respectively; temperatures ($^{\circ}\text{C}$) of determinations are shown.

found that the viscosity of the dry andesite melt at 15 kb and $1,350^{\circ}\text{C}$ was 850 poise, which was reduced to 45 poise by the addition of 4 wt.% water. This reduction is significant but small compared to the drop from 10^{11} to 10^6 poise seen for rhyolites, and the viscosity of hydrous rhyolite is still very much greater than for andesite. No experimental data have been published for hydrous basaltic melt but some theoretically deduced curves are shown in Fig. 3.

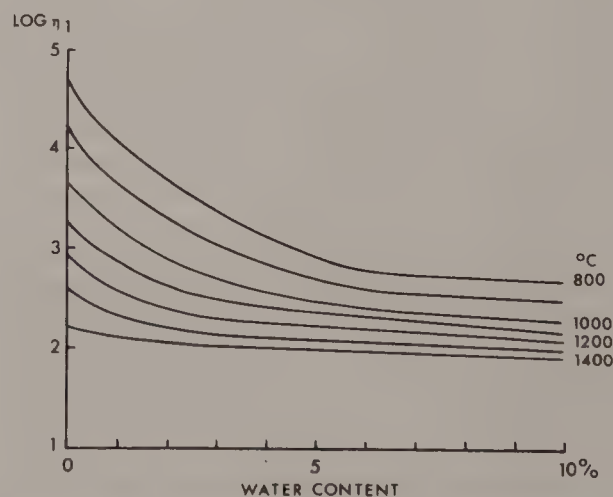


FIGURE 3. Viscosity (poises) as a function of water content (wt.%) at various temperatures ($^{\circ}\text{C}$) for a melt of basaltic composition. (After Murase, 1962.)

Lowering of liquidus and solidus temperatures. Compositions within the system SiO_2 – KAlSi_3O_8 – $\text{NaAlSi}_3\text{O}_8$ (the “granite system”) approximate closely the compositions of many rhyolites and granites and data from it can be used as an approximation to the effect of water on analogous rock melts. Figure 4 is a P – T projection of phase transitions occurring within the granite system. Of special interest in this discussion is curve II, which shows the lowering of the minimum melting point with increasing $P_{\text{H}_2\text{O}}$. The temperature of the ternary minimum in the dry system is ca. 960°C , and there is a rapid drop in the temperature to 720°C at 1 kb $P_{\text{H}_2\text{O}}$; from this pressure to 10 kb the temperature is lowered less quickly and is 615°C at 10 kb $P_{\text{H}_2\text{O}}$. From 0 to ca. 3.6 kb $P_{\text{H}_2\text{O}}$ the lowest melting point in the system is a ternary minimum; at about this pressure, two feldspar fields instead of one are stable at liquidus temperatures, and the ternary minimum becomes a ternary eutectic. The effect of $P_{\text{H}_2\text{O}}$ on the composition of the ternary minimum and ternary eutectic can be seen in Fig. 5: the

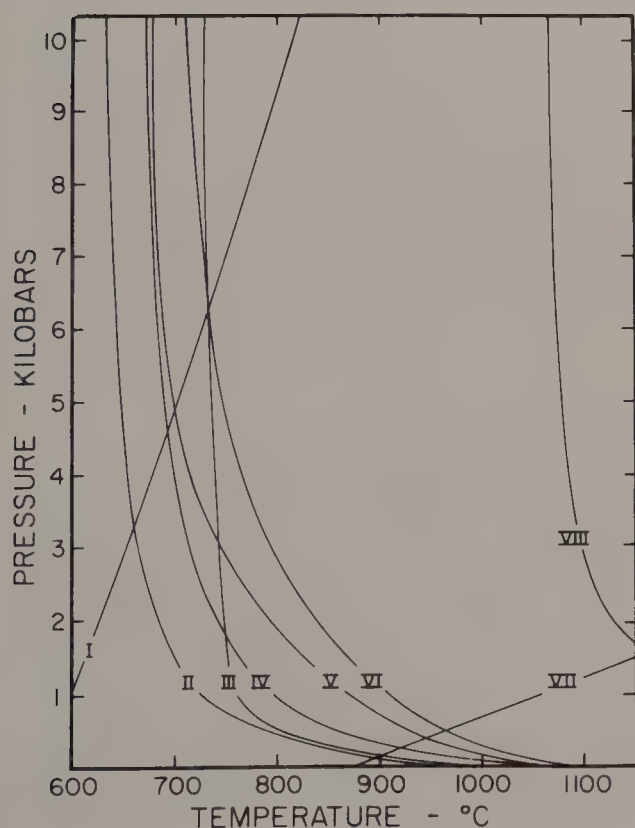


FIGURE 4. P – T projection of the beginning of melting in the granite system, other simpler systems, and two silica transformations. I, high–low quartz transformation; II, albite–orthoclase–silica–water; III, orthoclase–silica–water; IV, albite–silica–water; V, albite–orthoclase–water; VI, albite–water; VII, tridymite–quartz transformation; VIII, silica–water. (After Luth et al., 1964.)

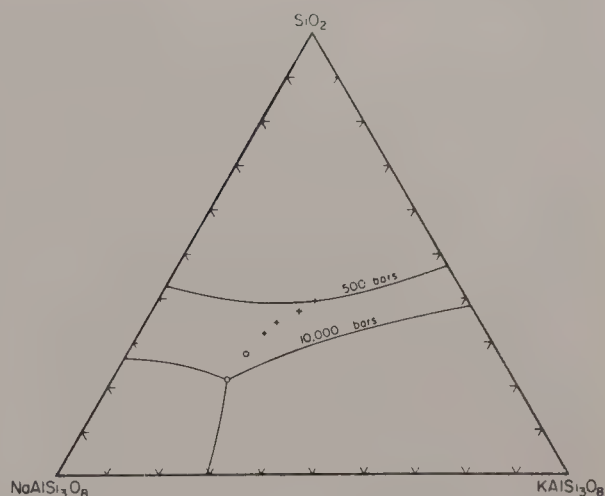


FIGURE 5. Diagram showing the progressive shift of the quaternary isobaric minimum and the quaternary isobaric eutectic in the granite system with increasing values of $P_{\text{H}_2\text{O}}$. Crosses indicate isobaric minimum; circles indicate isobaric eutectic; field boundaries are shown at 0.5 and 10 kb. (From Luth et al., 1964.)

anhydrous projection of the composition of the minimum at 500 bars $P_{\text{H}_2\text{O}}$ is at 31% KAlSi_3O_8 , 30% $\text{NaAlSi}_3\text{O}_8$, 39% SiO_2 in contrast to 21% KAlSi_3O_8 , 56% $\text{NaAlSi}_3\text{O}_8$, 23% SiO_2 at 10 kb $P_{\text{H}_2\text{O}}$.

No simple synthetic system is analogous to basalts, and it is probably for this reason that considerable data are available for natural basalt–water systems. The P – T projection of a natural olivine tholeiite–water system is shown in Fig. 6. Special points to be noted on this diagram are the lowering of the liquidus and beginning of melting temperatures, and the increased thermal stability of amphibole with increasing $P_{\text{H}_2\text{O}}$.

Effects of water on the mineralogy of rocks. The most obvious effect of water in rock melts on the subsequent mineralogy of the rock is an inducement for the crystallization of hydrous minerals. The two most common classes of hydrous minerals found in igneous rocks are amphibole and mica; the former may be abundant in a wide range of rock types, while mica is more restricted in distribution.

An indirect effect of high $P_{\text{H}_2\text{O}}$ is the possible precipitation of garnet in melts of basaltic composition. Work in the diopside–forsterite–anorthite system at 15 kb $P_{\text{H}_2\text{O}}$ has shown that garnet is the primary phase in part of this system. In this case the crystallization of garnet is not dependent on the presence of water in the melt, since garnets can form by a solid-state reaction in powdered basalt when subjected to a dry pressure of 15 kb. However at 15 kb $P_{\text{H}_2\text{O}}$

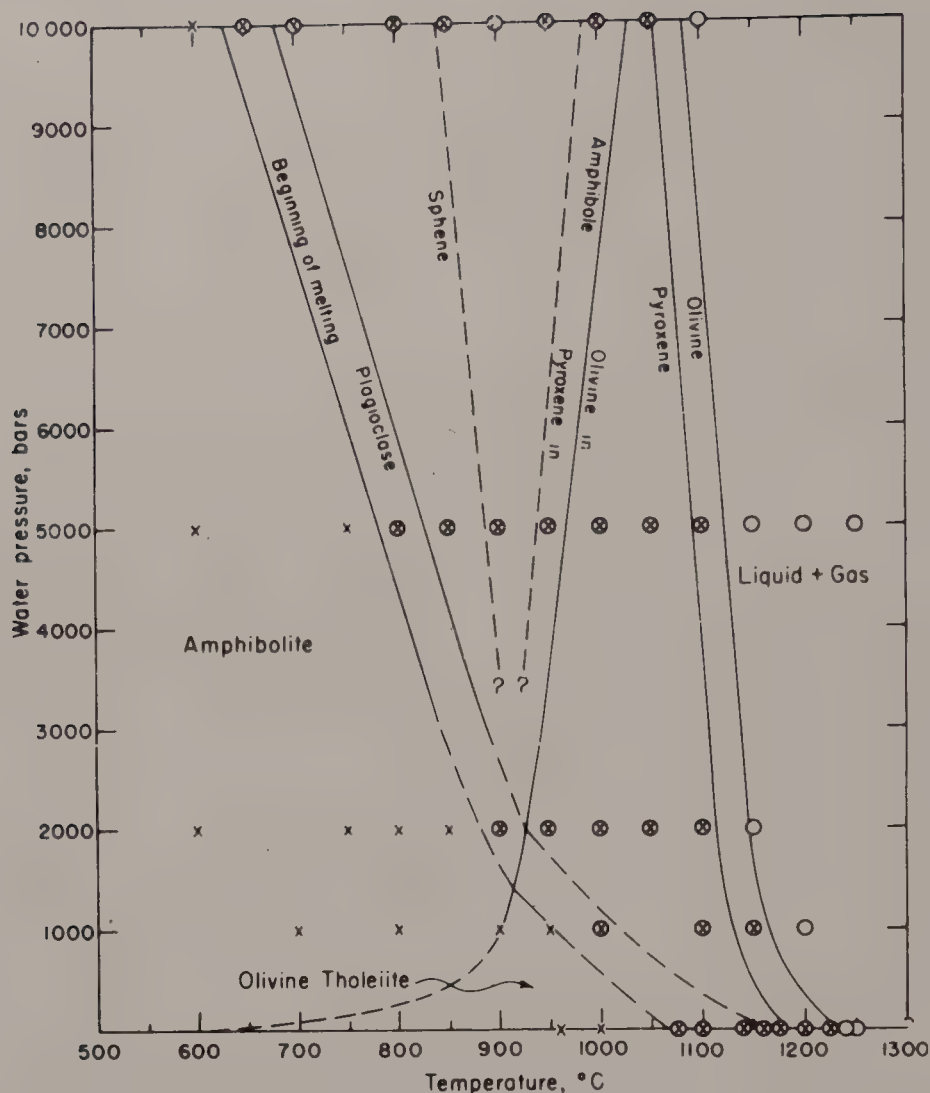


FIGURE 6. Projection of a natural olivine tholeiite–water system; mineral is stable on side of curve along which its name is printed. (From Yoder and Tilley, 1962.)

the temperature of the beginning of melting has been lowered until garnet and melt are compatible.

Effect of water on the partial pressure of oxygen of rock melts. A phase equilibria study of the system $\text{MgO-FeO-Fe}_2\text{O}_3\text{-SiO}_2$ (Osborn, 1959) has shown that the composition of successive liquids resulting from fractional crystallization will be in equilibrium with an ever-decreasing P_{O_2} . P_{O_2} of a water-saturated melt is influenced by the dissociation of water and hence the rate of fall of P_{O_2} with temperature of a dry melt and an analogous wet melt will not be the same. Further studies on more complex systems indicate that the rate of fall of P_{O_2} with temperature is essentially the same as determined for composition in the four-component system. The extrapolation from the more complex synthetic melts to natural rock melts is not too great and it seems likely

that the relation between temperature and P_{O_2} found in synthetic melts will also hold for natural rock melts. Thus, the presence or absence of water will have a profound effect on the P_{O_2} of successive liquids produced by fractional crystallization. In basalts, where the iron oxide content is commonly of the order of 10%, the level of P_{O_2} can greatly influence the type and composition of the precipitating phases: thus, the compositions of successive liquids, in some measure at least, are dependent on the level of P_{O_2} .

D. L. HAMILTON

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X

XENOLITH

Inclusions in magmatic rocks fall into two broad categories: *cognate xenoliths* are genetically related to the magma in which they occur and *accidental xenoliths* have no direct relation to their host. The latter may be derived from all levels down to the source region of the magma, and some volcanic vents that erupt explosively contain fragments derived from the entire section through which the magma has risen. It may be difficult to decide whether some coarse-grained mafic rocks are xenoliths torn from the mantle or accumulations of crystals from the magmatic liquid, because, with increasing depth, the composition of the lower crust and mantle become more like that of basic igneous rocks.

Cognate xenoliths

Cognate xenoliths have a variety of origins. Some are simply coarse-grained rocks formed by slow crystallization at depth, and their chemical composition may be very close to that of the lavas that bring them to the surface. Others are fragments of the lower crust or mantle and represent the original material from which magmatic liquids may be derived by partial melting. Still others are composed of crystals precipitated from a cooling magma or residual crystals left after partial melting and extraction of a liquid. All these types may be caught up by a rising magma and brought to the surface in lava flows or shallow intrusions. A wide variety of xenoliths may be found in the same host (Fig. 1).

The proportions and compositions of minerals vary not only with their mode of origin but also with the temperature and depth at which they were last in equilibrium with each other or with a magmatic liquid. The minerals in xenoliths from deep sources may be unstable at shallow depths and either break down to new assemblages or react with their host liquid. Al-bearing minerals show these pressure effects most clearly. Plagioclase is stable to depths equivalent to ca. 8–10 kb. Below that level it ceases to be stable and alumina enters pyroxene or spinel in increasing amounts until, at pressures of ca. 25–30 kb, garnet is a stable

phase and takes up most of the alumina of the rock. The exact pressures at which these changes take place depend on the bulk composition of the rock. Another important effect of pressure is the increased stability of Ca-poor pyroxene relative to olivine. Owing to the higher silica content of pyroxene, this feature may account for the observed association of xenoliths containing Ca-poor pyroxene (lherzolites, harzburgites, and websterites) with host lavas having low silica contents (Wyllie, 1967).

Silica-saturated magmas, such as tholeiitic



FIGURE 1. Ultramafic xenoliths in basaltic lava. The light-colored fragments are dunite with small amounts of spinel; the large fragment with dark grains is a lherzolite (olivine, clinopyroxene, enstatite, spinel).

basalts and andesites rarely contain cognate xenoliths other than gabbros having bulk compositions and mineral assemblages similar to or only moderately less siliceous than their host. Ultramafic cognate xenoliths are almost unknown in andesites, dacites, rhyolites, or their intrusive equivalents. But more silica-deficient magmas commonly contain, in addition to gabbroic inclusions, ultramafic rocks, such as dunite and spinel peridotites. Highly under-saturated basanites and lamprophyres may contain all the foregoing types as well as garnet- and enstatite-bearing rocks.

Accidental xenoliths

A wide variety of sedimentary, metamorphic, and older igneous rocks of the subvolcanic basement have been reported as accidental xenoliths in lavas and intrusive rocks, but many have undergone such severe alteration that their origin is obscure. Differential thermal expansion and heating of volatile components causes most rocks to disintegrate rapidly into small fragments. Metamorphic effects may produce pronounced mineralogical and textural changes, especially in xenoliths engulfed in large plutonic bodies or high-temperature lavas. Shale, slate, and mica schist have a wide range of alteration from minor hydrothermal discoloration to complete fusion. Many are reduced to porous white nodules of cordierite, sillimanite, tridymite, and glass. Rocks of more extreme compositions, such as limestone or quartz sandstone, are more resistant to melting. Sandstone caught up in high-temperature magmas may be partially fused to buchite, but a more common effect is recrystallization or reaction depending on whether the magma is saturated with silica. Thermally metamorphosed limestone and dolomite have been found among the lavas and pyroclastic ejecta of a number of volcanoes, notably Vesuvius, where the rocks are altered to a variety of calc-silicate minerals such as vesuvianite (idocrase), zoisite, grossular, wollastonite, diopside, and andalusite.

Quartz grains from metamorphic inclusions in lavas can often be recognized if they have preserved undulatory extinction or sutured grain boundaries. In basic lavas, quartz xenocrysts may persist as clear individual grains with well-rounded outlines and coronas of pyroxene; quartz in siliceous lavas tends to be deeply corroded, but this feature is not diagnostic of xenocrysts, because it is observed in phenocrysts as well. K-feldspar may be converted to sanidine; biotite and hornblende breakdown to pyroxene and magnetite.

In general, the rules derived by Bowen (1928) for assimilation of xenoliths will apply wherever



FIGURE 2. Xenolith of felsic gneiss in gabbro surrounded by a zone representing the products of partial melting; Skaergaard intrusion, E Greenland.

equilibrium is approached. A mineral immersed in a silicate liquid will melt or dissolve if the liquid is undersaturated with that mineral, the heat of fusion being derived from crystallization of other minerals with which the liquid is saturated at that stage of crystallization (Fig. 2). Xenocryst minerals earlier in the sequence of crystallization remain in the liquid without being dissolved, but the magma may react with them and convert them to compositions in equilibrium with the liquid at that stage. In both cases, the main effect of assimilation of xenoliths of most common rock types is to change the proportion but not the nature of minerals crystallized from the magma. The overall character of the liquid is not changed from what would be produced by normal crystal fractionation without the addition of xenoliths, but the proportions of liquids and crystalline assemblages of various compositions may be greatly changed. Trace-element and isotopic ratios, particularly those of Sr, Nd, and O, may be altered, however, and are normally the most useful means of detecting contamination of magmatic liquids in which xenoliths have been assimilated (McBirney, 1979).

Other terms

Hypoxenolith—derived from a source farther than the adjacent country rocks.

Raft—a rock fragment drifting freely in a magma.

Skialith—vague remnant of much changed country rocks: sometimes restricted to use in relation to the progress of feldspathization or granitization.

Xenocryst (chadacryst, cadacryst)—a crystal resembling a phenocryst that is foreign to the body of rock in which it occurs; the term *disomatic* used to be applied to what is now called a xenolith.

ALEXANDER R. MCBIRNEY

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Cross-references: *Assimilation*; *Buchite*; *Enclave*; *Lamprophyre*; *Mafic and ultramafic inclusions in igneous rocks*; *Magma*.

Z

ZEOLITE FACIES

The *zeolite facies* is a very low-grade mineral facies characterized by the presence of zeolites in rocks of appropriate composition. It includes products of the lowest grades of metamorphism in the sense that some of the mineral assemblages result from reconstitution of parent rocks in response to new conditions. These changes may result from burial metamorphism in thick sedimentary or volcanic piles, or from hydrothermal metamorphism on land or in the ocean crust. On the other hand, some zeolitic assemblages, such as those formed in alkaline lake beds, are strictly diagenetic. On casual examination many zeolite facies rocks appear to be little altered and they normally lack any trace of schistosity. To encompass burial metamorphic, hydrothermal, and diagenetic assemblages, the zeolite facies is better regarded as a broadly based mineral facies than as a metamorphic facies.

Occurrence of zeolites

In igneous rocks. Many of the finest specimens of zeolite come from amygdaloidal, scoriaceous, or brecciated basalts and dolerites, but these minerals are also known from zones of alteration and hydration in most other igneous rocks, including granites in which laumontite, heulandite, and stilbite are occasionally found in fractures and cavities, or partially replacing feldspar. In typical volcanic piles in Iceland and Northern Ireland individual zeolite species tend to occur in zones that transect volcanic stratigraphy and postdate the initial cooling (references in Coombs, 1971). Zeolites including analcime, natrolite, laumontite, thomsonite, chabazite, and stilbite are also found in basalts affected by ocean-floor metamorphism in hydrothermal systems at spreading ridges. On the other hand, historic lava flows in general do not contain zeolites. It should be noted that in the absence of free SiO_2 and at high $P_{\text{H}_2\text{O}}$, analcime might occur in certain alkalic rocks at magmatic temperatures; such occurrences are excluded from the zeolite facies by the definition given below.

In active geothermal fields. Zeolites are well known in many active geothermal fields and

their zonal distribution in relation to temperature has been revealed by drilling; examples include Yellowstone (Wyoming, U.S.A.), Wairakei (New Zealand), Onikobe (Japan), and Pauzhetka (Kamchatka, U.S.S.R.). In order of increasing temperature, a typical sequence of Ca-bearing zeolites is mordenite, heulandite or clinoptilolite, laumontite, sometimes yugawaralite, wairakite. In basaltic host rocks in Iceland, chabazite is reported at temperatures $< \text{ca. } 70^\circ\text{C}$, mesolite-scolecite $70\text{--}100^\circ$, heulandite and stilbite $70\text{--}150^\circ$, mordenite $80\text{--}230^\circ$, laumontite $100\text{--}230^\circ$ and wairakite $150\text{--}300^\circ$ (Kristmannsdóttir and Tómasson, 1978). The non-zeolite Ca-aluminosilicates prehnite and epidote sometimes appear in the higher-temperature sections of such drill holes. The generalized sequence is one of decreasing hydration with increasing temperature. Similarly, the Na-zeolite analcime coexisting with quartz gives way at higher temperatures to albite. At Sespe Hot Springs (California, U.S.A.) laumontite is reported to be precipitating with quartz on surface boulders lining a creek bed at temperatures of 89°C to 43°C (McCulloh et al., 1981). Other zeolites such as chabazite, phillipsite, and natrolite have long been known to crystallize in hydrothermal systems at similar low temperatures, e.g., in the masonry of ancient Roman baths.

In near-surface sediments. Various zeolites, notably analcime, clinoptilolite, phillipsite, chabazite, erionite, and mordenite are widespread and often voluminous products of diagenesis in alkaline lake beds (Sheppard, 1971). Silicic volcanic glass is a common precursor for such zeolite deposits, but analcime also occurs in non-tuffaceous non-marine sediments and is interpreted as formed in alkaline saline lakes either by direct precipitation, or by reaction between clays and brines. With aging, analcime, clinoptilolite, or feldspars tend to form at the expense of the other zeolites listed. Similarly, phillipsite is widespread in modern deep-sea sediments, whereas clinoptilolite is abundant in tuffaceous deep-sea sediments of Tertiary age. Zeolites other than analcime, heulandite or clinoptilolite, and laumontite are rare in pre-Tertiary sedimentary rocks. The activity ratios of alkali ions to hydrogen ions and the activity of silica in the

aqueous phase are important chemical controls for the formation of the various zeolite assemblages, clay minerals, or feldspars during early diagenesis.

In thick sedimentary and burial metamorphic sequences. Zeolites occur in greater or lesser abundance in many very thick sequences of sedimentary rock, especially those rich in detrital plagioclase feldspar and/or volcanic debris. Thus, in the Taringatura and the neighboring Hokonui districts of southern New Zealand, a Lower Mesozoic sequence nearly 10 km thick (Murihiku Supergroup) consists mainly of andesitic to rhyolitic volcanic graywackes and siltstones, with numerous thin ash beds. Laumontite occurs throughout the section as a replacement product of glass, as a partial replacement product of plagioclase, and as a cement. Laumontite is particularly abundant towards the base, where associated relict plagioclase is albitized, but it is also abundant as a cement and replacement product of non-albitized plagioclase in some of the uppermost strata preserved (Boles and Coombs, 1977). Laumontite and heulandite may partially replace shell material of fossils. A typical mineral assemblage is quartz–albite–laumontite–“chlorite”–sphene (titanite) $\pm$ celadonite $\pm$ calcite, the chlorite tending to form mixed layer structures with illite. The 1:1 ordered chlorite–smectite interlayered structure corrensites has often been reported from similar rocks from other parts of the world. Apart from minor relict quartz and augite, many of the rocks are mineralogically almost entirely reconstituted, even though sedimentary textures are largely unmodified. Prehnite and pumpellyite occur in the lower part of the section and prehnite spasmodically replaces heulandite in higher parts of the section. Analcime is lacking in the lower 3–4 km of the section, although its former presence is shown by pseudomorphs. Members of the heulandite–clinoptilolite series, often co-existing with smectites, occur throughout. Diagenetic heulandite–clinoptilolite and analcime are particularly characteristic of ash beds containing fresh unalbitized plagioclase in the upper part of the Taringatura section, their formation being favored by high a_{SiO_2} , resulting from the former presence of high-silica glass. Documented mineral changes include heulandite or clinoptilolite to analcime, heulandite or clinoptilolite to laumontite plus quartz plus feldspar, analcime to albite, heulandite and laumontite to prehnite plus quartz. In spite of gross overlapping of mineral zones, the direction of progressive alteration towards the quartz–albite–laumontite–chlorite assemblage is clear. With advancing metamorphism elsewhere in southern New Zealand, laumontite

assemblages in turn give way to non-zeolitic assemblages containing prehnite, pumpellyite, and/or lawsonite with albite and chlorite; smectite and mixed-layer phyllosilicates disappear.

Generally similar sequences are documented from many parts of the world including Puerto Rico, the European Alps, New England (Australia), California and parts of Oregon (U.S.A.), and Kii Peninsula, Japan. In regions considered to have had a steeper thermal gradient, e.g., Tanzawa Mountains of Japan, a zone of wairakite-bearing assemblages may intervene between the laumontite and zeolite-free zones (Coombs, 1971, and references therein).

Definition of the zeolite facies

Turner (1958) proposed a “zeolitic facies” largely on the basis of the observations in southern New Zealand, “to cover only regionally developed zeolitic assemblages that largely replace the pre-existing rocks and conform to the mineralogical and chemical requirements of a metamorphic facies....” Laumontite–albite–quartz, rather than heulandite or analcime, with quartz, was taken to be characteristic of the facies. Largely to avoid questionable distinctions between metamorphism and diagenesis, and on the basis that mineral assemblages reflect the physiochemical conditions of their formation whether they be metamorphic, hydrothermal, or diagenetic, Coombs et al. (1959) defined the zeolite facies as a *mineral facies* “to include at least all those assemblages produced under physical conditions in which quartz–analcime, quartz–heulandite, and quartz–laumontite are commonly formed.” Winkler (1967) followed Turner in restricting the zeolite facies to metamorphic assemblages containing laumontite as the critical phase and more specifically to the facies represented by Fig. 1d: this was termed the “laumontite–prehnite–quartz” facies. Separate mordenite, heulandite, laumontite, and wairakite facies or subfacies have also been proposed. Most workers would probably accept a broadly based definition of the zeolite facies accommodating all such quartz–zeolite assemblages, some of which are not necessarily thermodynamically the most stable ones under their conditions of formation.

Analysis of mineral assemblages

Some possible suites of zeolite facies mineral assemblages that can coexist with quartz are illustrated in Fig. 1. Possible additional phases include analcime or albite, K-feldspar, celadon-

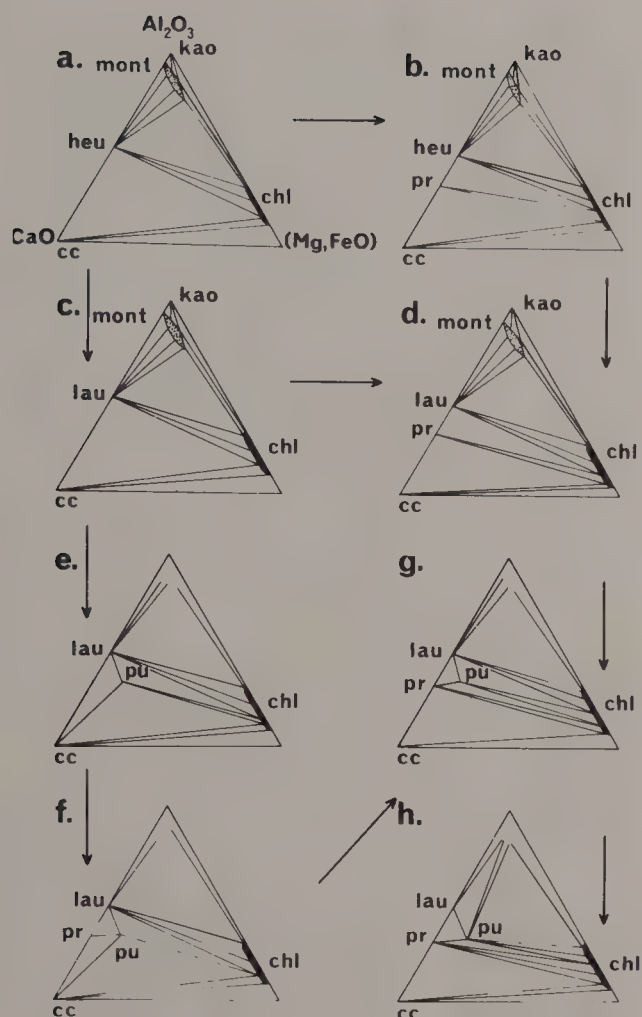


FIGURE 1. Some zeolite facies assemblages containing quartz and analcime or albite as additional phases: cc, calcite, chl, chlorite; heu, heulandite, kao, kaolinite; lau, laumontite, mont, montmorillonite, pr, prehnite, pu, pumpellyite. Further study of the appropriate aluminous clay minerals is required; the joins calcite–dolomite and dolomite–chlorite are also presumably stable. (From Coombs, 1971.)

ite, and mixed layer phyllosilicates. Arrows represent dehydration or decarbonation reactions and hence conventionally increasing metamorphic grade. The term “zeolite facies” is thus a collective name for all the suites of assemblages represented in Fig. 1, together with analogous facies including wairakite instead of laumontite, and a suite of very low-temperature diagenetic assemblages containing a variety of other zeolites. Under SiO_2 -deficient conditions other zeolites, especially thomsonite, may occur. It should be noted that prehnite often coexists with heulandite, and that both prehnite and pumpellyite may coexist with laumontite within assemblages here accepted as belonging to the zeolite facies. The assemblage calcite–chlorite–quartz–albite without other Ca-aluminosilicates is not diagnostic and may be

found in areas of zeolite, prehnite–pumpellyite, greenschist, and other low-grade facies.

Role of CO_2 and H_2O

It was first pointed out by E-an Zen that zeolite and clay–carbonate or greenschist facies assemblages can be related as a function of differing values of $\mu_{\text{H}_2\text{O}}$ and μ_{CO_2} under isothermal, isobaric conditions (Fig. 2). At low values of μ_{CO_2} , progressively decreasing $\mu_{\text{H}_2\text{O}}$ can result in assemblages that can be allotted in sequence to the zeolite, prehnite–pumpellyite, and greenschist facies. Although the diagram applies to isothermal, isobaric conditions, the same sequence can be predicted for progressively increasing temperature at low values of μ_{CO_2} . For higher values of μ_{CO_2} , each of these facies gives way to clay–carbonate or pyrophyllite–carbonate assemblages, and for suitably high values of μ_{CO_2} it is possible, either isothermally, or with increasing temperature, to pass directly from clay–carbonate to epidote-bearing greenschist facies assemblages without zeolites, prehnite, or pumpellyite, as has been recorded, for example, in the geothermal field of the Salton Sea, California, U.S.A.

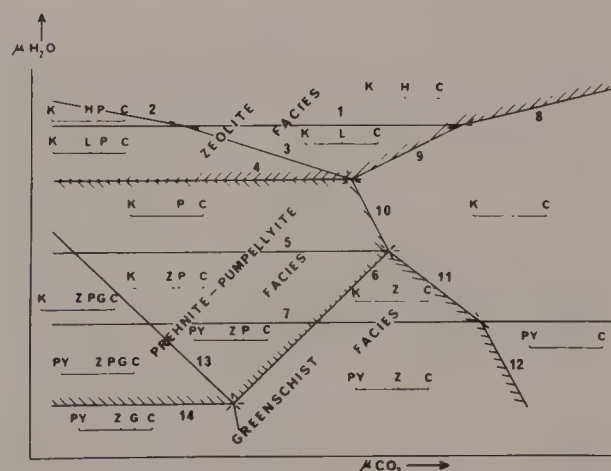


FIGURE 2. Schematic $\mu_{\text{H}_2\text{O}}-\mu_{\text{CO}_2}$ diagram for part of the system $\text{Al}_2\text{O}_3\text{--CaO--SiO}_2$ in the presence of excess quartz, at some arbitrary value of T and total P ; the position of the kaolinite–pyrophyllite reaction, relative to the other reactions shown, is uncertain. C, calcite, G, grossular; H, heulandite; K, kaolinite; L, laumontite; P, prehnite, PY, pyrophyllite; Z, zoisite.

In any divariant assemblage, two of these phases can coexist as indicated by adjacent symbols in the phase compatibility diagrams; e.g. in the uppermost field (above the univariant lines 2, 1, and 8) the phases kaolinite, heulandite, and calcite are stable where the ratio of $\text{Al}_2\text{O}_3\text{:CaO}$ is 1:0, 1:1, and 0:1, respectively. Kaolinite and heulandite coexist for finite ratios of $\text{Al}_2\text{O}_3\text{:Ca} > 1:1$; heulandite and calcite coexist for finite ratios $\text{Al}_2\text{O}_3\text{:Ca} < 1:1$. (From Coombs, 1971.)

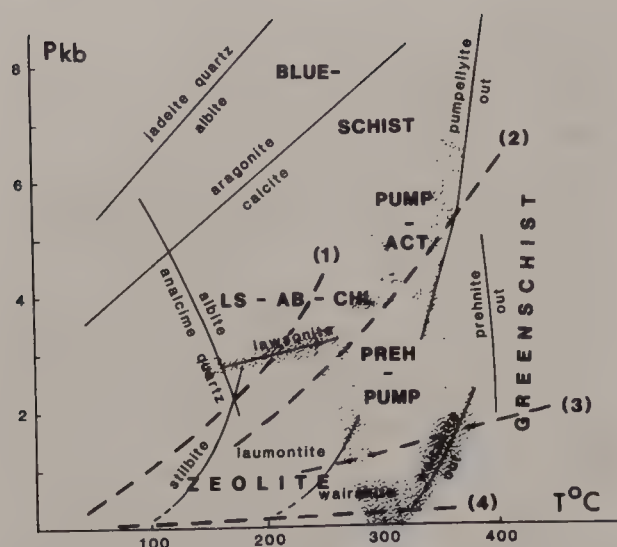


FIGURE 3. Petrogenetic grid showing univariant reactions critical to the zeolite and contiguous very low-grade mineral facies. Shading represents provisional positions of transition zones between zeolite, lawsonite-albite-chlorite, prehnite-pumpellyite, pumpellyite-actinolite, and greenschist facies: the dashed trajectories 1, 2, 3, 4 represent high-pressure, intermediate-pressure, low-pressure, and very low-pressure facies series. (Data from Coombs, 1971, Liou, 1971, Liou et al., 1983, and references therein.)

Stability relations

Early attempts to establish stability relations of critical zeolite facies minerals were largely frustrated by failure to establish reversible equilibria. Liou (1971) resolved the problem for a number of critical reactions by simultaneously holding charges of mineral reactants and products in 9:1 and 1:9 ratios at appropriate temperatures and pressures for periods long enough to produce unequivocal growth of seeds (products) at the expense of reactants. The equilibria portrayed in Fig. 3 apply to the conditions $P_{H_2O} = P_{TOTAL}$. Under conditions where P_{H_2O} is less than P_{TOTAL} , reaction temperatures are displaced to lower temperatures. Similarly, the equilibria are displaced to lower temperatures in the presence of solutes.

Petrogenetic grid for very low-grade mineral facies

At pressures in excess of ca. 3 kb and temperatures of ca. 200–250°C, the most characteristic zeolite facies mineral, laumontite, breaks down to lawsonite, quartz and H_2O and the critical assemblage for the lawsonite-albite-chlorite facies become stable. The field transition across this boundary (curve 1 of Fig. 3) exemplifies a high-pressure, low-temperature facies series, as in (1) the Franciscan rocks of California (U.S.A.), (2) parts of the Sanbagawa

terrane of Japan, and (3) the Maitai Group of southern New Zealand. Curve 2 represents the more normal case of an intermediate-pressure facies series. Analcime and heulandite or stilbite may survive from the early diagenetic stage, or may themselves be formed during burial metamorphism. They are followed by broad zones of laumontite, prehnite-pumpellyite facies, and pumpellyite-actinolite schist facies passing into greenschist facies. Curves 3 and 4 represent low-pressure facies series as in the Tanzawa Mountains (Japan) and many geothermal drill holes, respectively, with a narrower zone of laumontite, a zone of wairakite, and often no pumpellyite.

It must be emphasized that the controls over zeolite and associated mineral assemblages are complex. Broad limits are imposed by temperature and pressure, but ionic activity ratios in the associated fluid phase, (μ_{CO_2}), variations in P_{FLUID} relative to P_{TOTAL} in strata of varying permeability, and f_{O_2} in the case of Fe-bearing minerals are all of vital importance. Nucleation and kinetic effects may also be significant.

Reaction curves in Fig. 3 indicate maximum temperatures of stability of a number of key minerals. Where $P_{H_2O} < P_{TOTAL}$, or where the fluid phase is not pure H_2O , dehydration reaction temperatures will be lowered from those shown. Furthermore, stabilities are modified by solid-solution effects and key minerals may disappear from natural assemblages by more complex reactions than those shown. Although temperatures in natural hydrothermal systems show reasonable agreement with trajectory 4 in Fig. 3, the net effect is for natural facies boundaries to occur at rather lower temperatures than those indicated for the ideal end-member reactions portrayed. A low-pressure prehnite-actinolite facies has recently been recognized between the prehnite-pumpellyite or zeolite facies and the greenschist facies (e.g., Liou et al., 1987).

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